METHODS OF SINGLE STATION

AND LIMITED DATA

ANALYSIS AND FORECASTING

by

WALTER K. HENRY, Lt. Col. USAF (Ret.)

and

KENNETH C. BRUNDIDGE, Ph. D.

15 AUG 1985

AIR WEATHER SERVICE (MAC)

SCOTT AFB, ILLINOIS 62225

REPORT DOCUMENTATION PAGE

2. Report Date: 15 August 1985

3. Report Type Final

4. Title and Subtitle: Methods of Single Station and Limited Data Analysis and Forecasting

5. Funding Numbers: C F11623-84-C0047

6. Authors: Walter K. Henry and Kenneth C. Brundidge

7. Performing Organization Name and Address: Texas A&M University Research Foundation, College Station, Texas, 77843-3146

8. Performing Organization Report Number: N/A

9. Sponsoring Agency Name and Address: HQ Air Weather Service (AWS/DNT), Scott AFB, IL 62225-5008

10. Sponsoring Agency Report Number: N/A

11. Supplementary Notes: Fulfills contract F11623-84-C0047 between USAF/AWS and Texas A&M University Research Foundation.

12a. Distribution/Availability Statement: Approved for public release; distribution is unlimited

13. Abstract: A detailed discussion of "single-station weather analysis," from theory through practical application. Authors define single-station analysis (SSA) as "a procedure whereby the [weather] observations at a location are utilized to their fullest extent to provide information, either directly or by inference, about the meteorological conditions and the changes they are undergoing in the near vicinity of the given location." Discusses basic principles and theory, describes procedures for processing available data, and provides an example using real data. Discusses modifications of SSA technique in certain climatological regimes and describes some statistical techniques for SSA of specified phenomena.

14. Subject Terms: WEATHER, WEATHER FORECASTING, WEATHER ANALYSIS, WEATHER PHENOMENA, METEOROLOGY, SINGLE STATION, SINGLE-STATION ANALYSIS,

15. Number of Pages: 205

17. Security Classification of Report: UNCLASSIFIED

18. Security Classification of this Page: UNCLASSIFIED

19. Security Classification of Abstract: UNCLASSIFIED

20. Limitation of Abstract: UL

Standard Form 298

PREFACE

This report was written because the authors have received a great number of individual requests for this type of information. The authors are appreciative of the comments and suggestions of Dr. Dusan Djuric, Dr. Guy Franceschini, and Dr. Harry Thompson, all of Texas A&M University. They also wish to acknowledge the contributions of Major Robert Korose and other personnel of the Air Weather Service who contributed suggestions.

A special thanks must go to Miss Jennifer L. Richardson who typed the many revised drafts and who drew the diagrams.

CONTENTS

LIST OF ILLUSTRATIONS

LIST OF TABLES

CHAPTER I

INTRODUCTION

Prior to the development of any type of electronic communications, weather forecasts were of necessity based on very minimal data. Indeed, for the most part, input data were limited to those that could be observed at a single location. This was true for most land stations as well as all ships at sea. Synoptic meteorology did not develop until after telegraph networks became operational so that meteorological observations could be collected and analyzed in real time.

By the beginning of World War II, forecasts based on synoptic weather maps for large parts of the world were being made. However, the war quickly changed this situation as the combatants attempted to deprive their opponents of vital weather information. Once again, ships at sea, island stations and isolated combat groups found themselves without a data network, or had data from only a limited area. As a consequence, it became necessary to develop a technique that would permit the forecaster to make optimum use of the information available to him from a combination of rawinsonde and surface data for his station. The technique became known as "single station analysis" (SSA). Early publications dealing with the technique were by Rossby, Oliver and Boyden (1942), Boyden et al. (1942) and Jones and staff members (1943). Oliver and Oliver (1945) combined and organized the material and it was published in the Handbook of Meteorology.

Today, the meteorologist has the benefit of a world-wide network of surface and upper-air stations. Many have at their finger tips radar and satellite observations, computer-prepared analyses and forecasts, and statistical aids such as conditional climatology tables and model-output statistics (MOS). However, for some, the situation may be little different than it was during World War II. The weather data may be discontinued at any time by hostile action, communication network failure or by the requirement of the mission. Therefore, it seems reasonable to review the fundamental concepts of SSA and to update the procedures where possible. Furthermore, a knowledge of SSA procedures will help to improve any analysis--many features that are revealed may be overlooked in general synoptic analysis.

As the name implies, single station analysis is a procedure whereby the observations at a location are utilized to their fullest extent to provide

information, either directly or by inference, about the meteorological conditions and the changes they are undergoing in the near vicinity of the given location. All of the principles of analysis in standard practice still apply in SSA. In fact, strict and accurate spatial and temporal continuity become critical. To apply the SSA procedures, the forecaster must have a thorough understanding of the three-dimensional structure of synoptic systems, knowledge of seasonal climatology, and knowledge of local conditions that impact upon the meteorological events at his station. These points will be reiterated and expanded upon in the following chapters.

One can visualize a variety of situations in which a forecaster might find himself that vary in the amount of information available to him and the circumstances in which he is operating. His approach to the forecast problem must vary accordingly. Space will not permit consideration of every possible situation that might arise; therefore, in the chapters that follow we will concentrate on several examples which should point the way for most any circumstance. In Chapter II, we deal with the forecaster who is completely isolated and with little or no instrumentation. Chapter III will cover the classical SSA situation in which surface observations with an upper-air sounding are available and in Chapter IV the data base will be expanded to include information from three upper-air stations. In none of the cases will it be assumed that the forecaster has access to radar or satellite observations, or any computer products during the period of his isolation. Where calculations are involved, it is assumed that a hand-held calculator will be available. Each of these chapters will provide a discussion of the basic principles and theory applicable to the situation (PART A), procedures will be described to process the data (PART B) and an example using real data will be presented (PART C). If the reader does not wish to examine the theory of why certain procedures are used, he can skip PART A, and do PART B and PART C on faith.

The final chapters of the text will deal with special topics such as modifications of the SSA technique that must be considered for certain climatological regimes (Chapter V), and a brief discussion of some of the statistical techniques that have been developed for single-station forecasting of certain phenomena (Chapter VI).

To do SSA, certain materials and equipment are needed that are not routinely used in an AWS/USAF weather station. A list of items to have ready in

the event that the forecaster needs to resort to SSA is in the Appendix, Part A.

Part B of the Appendix has a list of symbols and abbreviations used in this manual.

CHAPTER II

OBSERVATION, ANALYSIS AND FORECASTING USING MINIMAL INSTRUMENTATION

PART A THEORY AND BASIC PRINCIPLES

Observations Without Instruments. For centuries people survived without using weather instruments when they decided what the weather might be. Many a ship set sail after the "Master" walked around the deck to "sniff" the weather. Not all the "Masters" made correct forecasts and not all ships made harbor safely. Aristotle wrote "Meterologica" about 300 B.C. and he knew many of the correct meteorological processes and he did not have th instruments that are in a weather station today.

Observations of many atmospheric conditions can be made without instruments. A list would include: clouds, type of precipitation, visibility, wind, stability of atmosphere, and a general sense of temperature and humidity. There will be a certain degree of precision missing, but many aspects of the atmosphere can be determined. A forecaster is never without some information that can be utilized.

The theory concerning observations without instruments is limited and no further discussion is needed in this part. The Federal Meteorological Handbook No. 1 has instructions for taking the visual part of the observation.

PART B PRACTICAL PROCEDURES

1. Detail the Observations. It is necessary that the observer (forecaster) be able to make the visual part of a standard observation and go beyond, to describe conditions even better. For example, a sky condition of "2000 scattered" could be a few flat cumulus clouds floating at random across the celestrial dome or could be an approaching line of thunderstorms. A few altocumulus standing lenticularis north (ACSL N) would indicate a different situation than just 10,000 scattered, or few altocumulus.

2. Background Knowledge. Assume a wartime situation in which the enemy territory represents a silent area as far as weather information is concerned. The forecaster may be alone or part of a combat team that the forecaster is to support in the area. First, consider a case in which the forecaster has no instrumentation but does have a watch and can determine directions. He should have a detailed map of the area. If he is to be

effective, there is a vast amount of information the forecaster must have at the moment he starts his mission. This information is as follows:

a. Climatology. The forecaster needs to know the climatology of the area for the given season. This should include familiarity with the weather events which are common. A knowledge of extremes will prevent an observation or forecast that is almost impossible. Also needed is knowledge of the diurnal variations of the meteorological variables for the given area and season. Reference should be made to the discussion by Jenkins (1945) for a description of the variability of the standard variables, not only diurnally, but also as a consequence of such factors as land vs. ocean locales or polar vs. tropical regions. Climatic classifications such as Köppen can give helpful information.

b. Landforms. The forecaster needs a knowledge of the topography and geography of the area. One needs to know of the presence of rivers, lakes and significant mountains and the influence these features may have on the weather of the area. World aeronautical maps are available and show the required information in sufficient detail.

c. Structure. The forecaster should have a thorough knowledge of the three-dimensional structure of atmospheric pressure systems and the sequence of weather events normally associated with moving systems. This is a crucial factor for the ability to make a forecast on the basis only of what the forecaster can observe at his location.

d. Currency. The forecaster should have the latest available information on, or estimation of, the weather conditions in the area when he departs to go to the new area. If it is available, a 12 h forecast for the area would be very useful.

3. <u>The Observations</u>. Before he can make a forecast, the isolated forecaster must first make a series of observations of the pertinent weather elements in order to establish trends. All observations should be recorded, and it is often useful to plot them to help determine trends. Without instrumentation this is a challenging task which demands considerable experience. The Federal Meteorological Handbook #1 is the official guide for making surface observations and should be reviewed. A listing of the elements of the observation and some comments on each are given below.

a. Clouds. The experienced observer knows the cloud types, their diurnal variations and their implications for producing weather or modifying the diurnal temperature vaiiation. He knows that halos are associated with cirrostratus and that coronas occur with altostratus. Furthermore, he is aware of the sequence of cloud types typically associated with moving systems, such as a wave cyclone. Figures II-B-1a and II-B-1b show the standard wave model. The orientation may be different and the dashed arrows showing north indicate the possible change in orientation. The double line arrow indicates the movement of the wave and the lines of arrows show the upper air wind direction.

b. Atmospheric Phenomena. This item covers the entire range of precipitation types and obstructions to visibility. Experience is required to recognize them. St. .c on his AM radio may indicate to the observer that lightning is occurring at some distance even when it is not directly observable.

c. Visibility. Again, experience plays an important role in making this estimation. However, prior knowledge of the distance to observable landmarks can be useful. The turbidity of the air can be a factor--this may be indicated by the sky color.

d. Pressure. There is no non-instrumental technique for obtaining the pressure. However, estimates of pressure change can be made from one's knowledge of the diurnal pressure variation for t.e given locale and estimation of the pressure systems moving through the area. A simple indicator may be made by placing a thin rubber covering over a jar and fixing it with a rubber band. As the pressure changes, the rubber will sink or extend from the top of the jar.

e. Temperature and Humidity. Knowledge of the local climatology and the diurnal variations of these elements must be utilized along with estimation based on one's personal comfort. Certain temperatures such as freezing can be determined. Snow will "crunch" when walked on at temperatures below 0° F. Advective changes in these properties generally can be sensed, thus giving a clue to air mass changes that may be in progress. However, care must be taken to distinguish between real cold temperature and apparent cold temperature as a

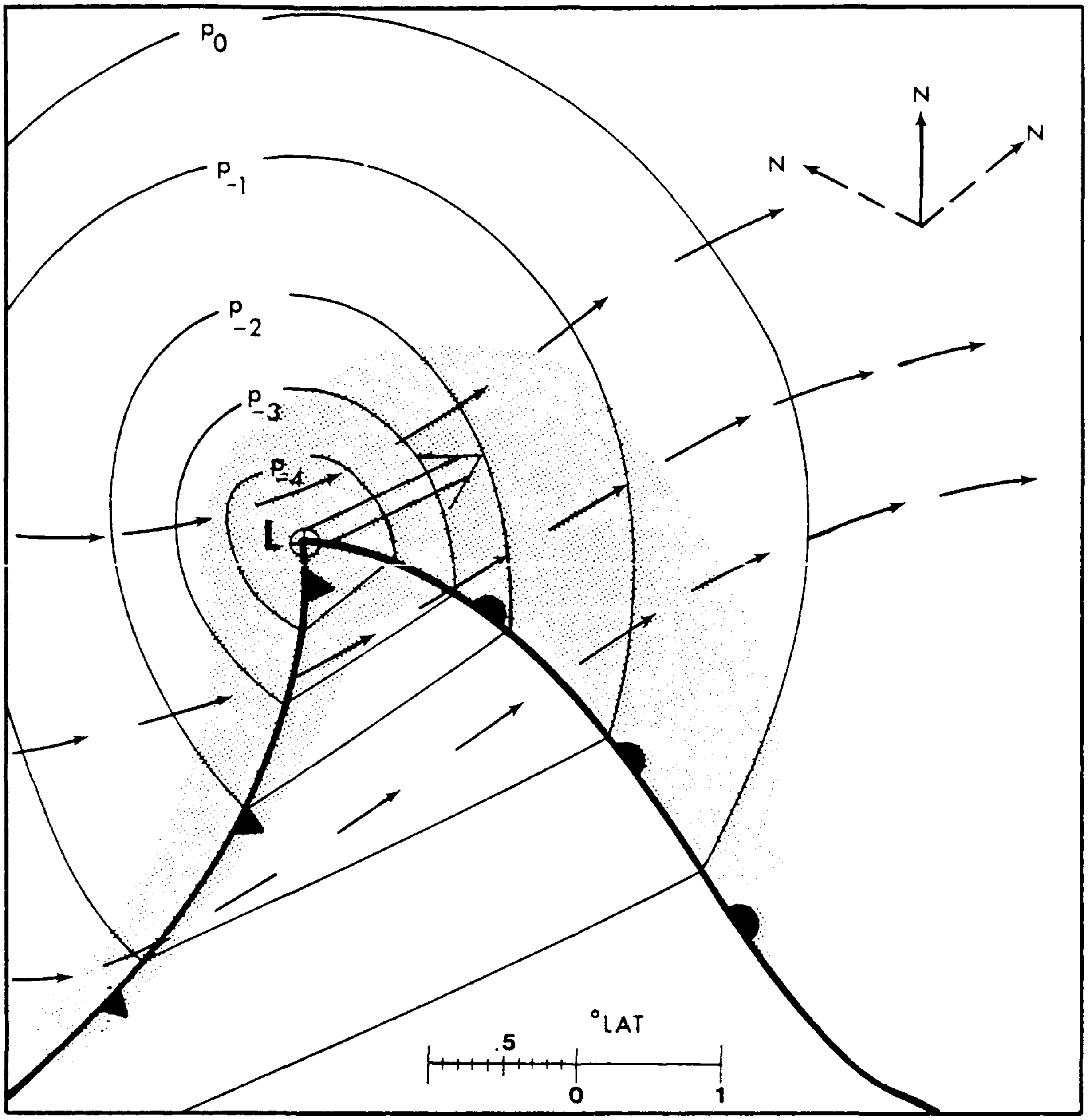


FIG. II-B-1a Sketch of open wave. The low, with the warm and cold fronts may have different orientations as shown by the different directions of north in the upper right hand corner. The double arrow indicates the direction of movement of the wave. The single arrows show winds aloft. The stippled area represents precipitation. The wave usually will progress to the occlusion stage.

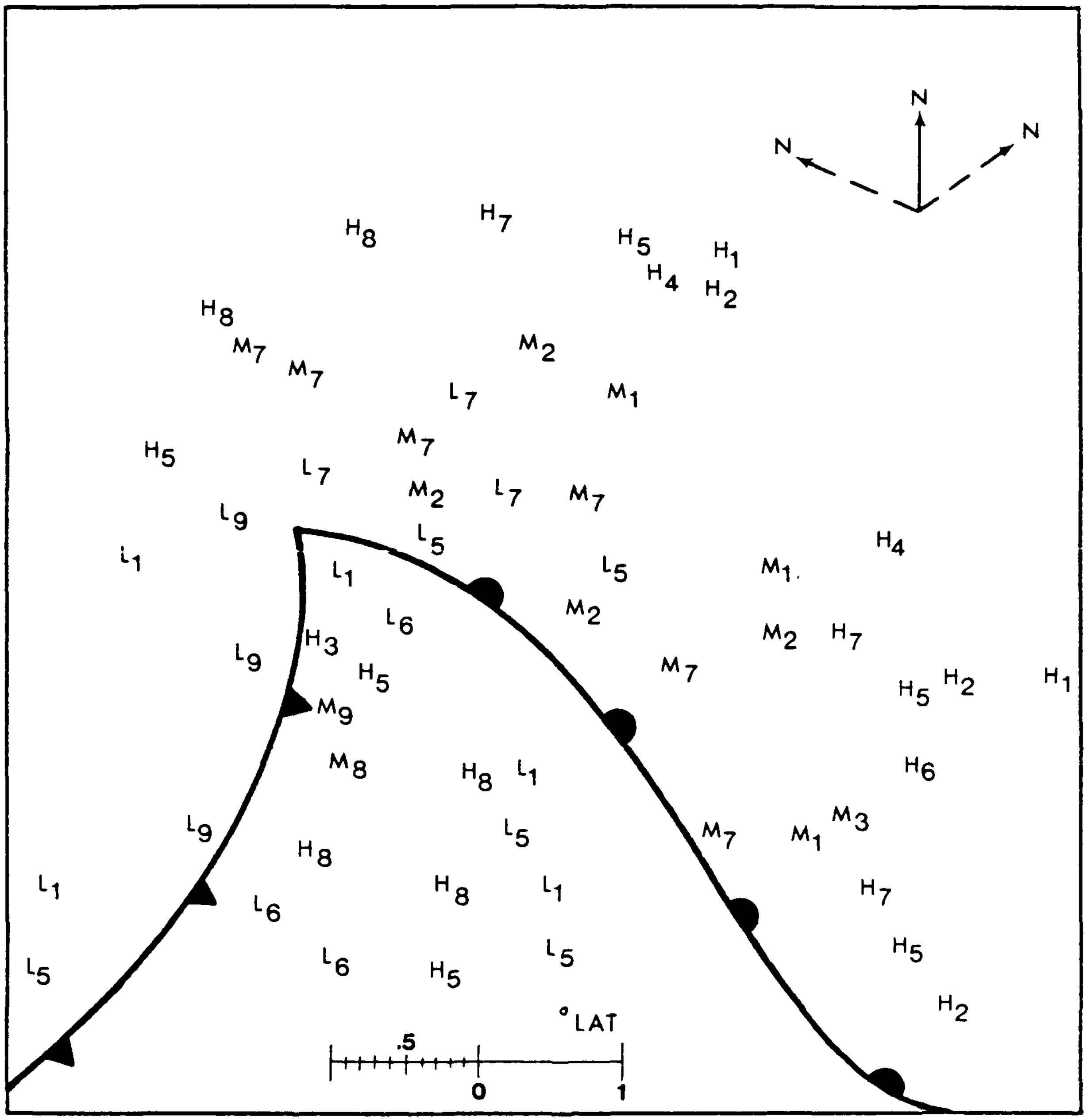


FIG. II-B-1b Cloud types with wave. The cloud types which usually occur in relation to the frontal wave are indicated. If the lower clouds form an overcast then the higher clouds will not be observed, but they will be there. See Federal Meteorological Handbook No. 1, Surface Observations for meaning of cloud types.

consequence of wind chill. Also the body's response to direct sunlight can be misleading for temperature estimation. The body does not sense humidity very well; however, with low RH (below 20%) the lips get dry and may crack, and the nostrils may feel "crimped" inside. Whether or not dew or frost forms at night gives some clue to the relative humidity. Fog or drizzle indicates a RH above 95%. A glow around street lights at night also indicates high humidity. It is possible to estimate the dew point spread aloft from visual observations at the surface. A "rule of thumb" for doing this using observations of low and middle clouds is given in the Appendix, Part C. A more elaborate method involving observations of cloud types and amounts, past and present weather, has been developed by Ball et al. (1965). The method uses decision trees for the 850, 700, 500, and 400 mb levels developed for the winter months over the United States. These trees with additional information are reproduced in the Appendix, Part D.

f. Wind. The wind direction at the surface can be obtained by observing the motion of smoke, dust, leaves, etc. The wind speed can be estimated from the Beaufort Scale which appears as Part E in the Appendix. Winds aloft can be estimated from cloud motion in a variety of ways. For example, puddles of water can be used as a nephoscope to get the cloud motion; or, one can stand with his back to a building with known orientation and observe the cloud motion relative to the building. The estimation of the wind speed aloft first requires a determination or estimation of the cloud height in order to apply geometric principles to the calculation. Various techniques for making wind estimates based on cloud motion are further described in the Appendix, Part F.

g. Stability. Stability in the boundary layer is known to vary diurnally. The less stable conditions normally occurring during the daytime are indicated by convective cloud types and gustiness of the surface wind. However, even at night, strong winds produced by a large pressure gradient can cause mechanical mixing and a lapse rate near the adiabatic value. Whirlwinds or dust devils are an indication of unstable air in the boundary layer. Stable layers aloft can produce a definite top to a haze layer and cause convective clouds to flatten

out and spread in the direction of the wind. A ground-based inversion, such as the nocturnal radiation inversion, permits sound to be heard over greater distances. Much can be learned about the boundary layer stability and its variation with time by observing the behavior of smoke from fires or stacks. Part G of the Appendix illustrates this aspect.

4. Starting. During approximately the first 12 h the isolated forecaster has little for guidance except for whatever weather information he had at the beginning. It follows that he must rely mainly on persistence and use his own observations to indicate the modifications that should be applied. As noted above, a crucial factor will be his ability to recognize his position relative to the synoptic system producing the conditions in his area and to use his observations of the weather elements to indicate the movement of the system in relation to his position.

PART C AN EXAMPLE

1. Setting. The weather team has been dispatched into an area to take observations and forecast in support of limited ground forces. During the move into a location identified as Station #3, the observation kit was lost but a watch, compass, radio, map, paper and pencils were available. The team proceeded to the specified location and started operations without instruments. The location is about 38° N at an altitude of 887 ft. The topography map indicates that the area has low rolling hills. About half of the area is in crop production and the rest is pasture land. To the southeast about five miles is a small city which is mostly residential and has only a few tall smoke stacks. Small lakes are in the area but no large bodies of water or major orographic features are within 500 mi. The month is February and the forecaster has reviewed the climatology of the area which is summarized in Part H of the Appendix. The forecaster was current with the known synoptic conditions for the area for Day 1 (D1) at 0600 LST.* A stationary front was known to be about 5° Lat south of Station #3 at that time.

*For this section, the time is given in 24-h Local Standard Time (LST instead of GMT to let the reader know day and night to convey the concept that SSA is not dependent upon a time or location. In real conditions it would better to use GMT and draw maps at standard times.

2. Data. The team established their station observation site, and determined north and the distance to several points for visibility markers. The first observation was recorded at D1-1755 LST which was just after sunset. The observations are shown in Table II-C-1.

3. Interpretation of Data, D1-1755 LST to D2-0557 LST.**

   a. 1755 LST. In cold air mass below freezing--maybe cP for a first guess.

   b. 1755 LST. Lapse rate in lower layer is close to dry adiabatic--smoke rising and dispersing downwind.

   c. 1955 LST. No fog tonight. The middle overcast and lower clouds that are forming will prevent a radiation inversion. Temperature should not lower during the night.

   d. 1955 LST. The altocumulus clouds indicate that some turbulence exists at the 700 mb level.

   e. 2055 LST. Warm air advection aloft is occurring, (lower wind west, upper wind northwest).

   f. 2155 LST. The layer of middle clouds continued to lower. Using the moisture decision trees in Part C of the Appendix, the 850 mb $T-T_d$ is estimated to be 3° C and the same at 700 mb.

   g. 2256 LST. The 4500 ft layer of stratus is variable in amount and represents the base of the inversion with a dry lapse rate below the cloud layer. The air mass must be cPk.

   h. Wind. The west wind at 2356 LST does not fit with the rest of the data. It may be a local drainage or an error. The calm wind of the next hour may indicate a radiation inversion forming, or the center of a high pressure area. Since the temperature does not seem to be colder and the overcast is solid the high pressure center is the better estimate.

   i. z-t Diagram. From a plot of the clouds on a z-t diagram (a z-t diagram can be drawn on any sheet of paper) by 2200 LST the slope of

---

**The forecaster may be at the observation site, or at some location outside of area and be receiving the reports by radio. The analysis and forecast are presented with the view that no other information is available regardless of where the forecaster is located. A forecaster at the site would have a better "feel" and make a better analysis and forecast.

TABLE II-C-1

VISUAL OBSERVATIONS TAKEN AT STATION #3 D1 1755 TO D2 0557 LST

DAY	TIME (LST)	SKY	VSBY	PRESSURE (MB)	TEMP (°F)	T_D (°F)	WIND DIR (°)	WIND SPEED (KT)	ALT SETTING (IN)	SKY-DETAIL N_s	CC	$h_sh_sh_s$	N_s	CC	$h_sh_sh_s$		REMARKS
D1	1755	E120 OVC	7+		COLD		NW	E10		8	AS	120	2	AC	120		AC MVG SE / Smoke rising and
					FROZEN												drifting SE Lapse rate neutral
D1	1855	E100 OVC	7+		COLD		NW	E10		8	AS	100	2	AC	100		
D1	1955	E55 BKN 100/OVC	7+		COLD		NNW	E10		6	SC	55	8	AS	100	2 AC 100	
D1	2055	55 SCT E 100/OVC	7+		COLD		N	E5		4	SC	55	7	AS	100	3 AC 100	SC MVG E AS MVG SE
D1	2155	E90 OVC	7+		COLD		ENE	E2		7	AS	90	3	AC	90		
D1	2256	E45 BKN 90 OVC	7+		COLD		NNE	E2		6	SC	45	7	AS	90	3 AC 90	850 F-Td E 3°C 100 F-Td E 3°C
D1	2356	45 SCT E80 OVC	7+		E15		W	E2		4	SC	45	7	AS	80	3 AC 80	
D2	0054	E40 BKN	7+		COLD		CALM	E0		6	SC	40					
D2	0157	E35 BKN	7+		COLD		SE	E3		6	SC	35					
D2	0255	35 SCT	7+		COLD		SSE	E5		4	SC	35					SC MVG E
D2	0315	E35 OVC	7+				SSE	E5		10	SC	35					
D2	0358	E30 OVC	7+		COLD		SE	E5		10	SC	30					
D2	0455	25 SCT E40 OVC	7+		COLD		ESE	E10		3	SC	25	10	ST	40		SC MVG NE
D2	0557	E25 OVC	7+		COLD		SSE	E10		10	SC	25					

the base of the clouds can be established (Fig. II-C-1). Knowing of the warm advection, a warm front should be identified. The slope extrapolates to a passage at D2-1100 LST. The front is not causing precipitation because the winds above it are from the NW which are parallel to the front. The clouds above the front with NW winds are not thick.

j. Low Clouds. The stratocumulus clouds are below the front and the moisture is left from previous showers as the front moved southward as a cold front.

k. Clouds Lower. The continued lowering of the stratocumulus clouds from 2256 to 0557 LST indicate that the warm front is approaching.

l. Strong Wind. The strengthening surface winds from the southeast quadrant at 0557 LST indicate that the warm front is approaching.

m. Surface Map. A surface map can be constructed at D2-6000 LST. Plot the data on the map (see Fig. II-C-2) and draw the standard frontal wave model. The front is about 5 h from passage, maybe 30-40 mi. The high to the northeast was placed there to accommodate the NW winds of the D1-1800 LST observations and the calm of the D2-0000 LST observation.

4. <u>Weather Briefing at D2-0730 LST</u>.

a. Front. The warm front is approaching and should pass by D2-1100 LST.

b. Clouds. No low clouds after the frontal passage.

c. Temperature. The temperature will rise to above freezing, 40° to 45° F, after frontal passage with some sunshine. New air mass probably mP. Warm front passage will cause warming, now below freezing. 42.7° F is the mean maximum temperature according to the climatology (Part H of Appendix).

d. Winds. Winds will be southerly rest of day.

e. Outlook. Cold front will pass in the next 12 to 24 h. (This is based on the frontal wave model).

f. Showers. Low clouds will form prior to cold front passage with 30% chance of showers (based on the frontal model).

5. <u>Analysis of Observations, D2-0600 LST to D2-1800 LST</u>. With the arrival of daylight the weather team found the missing observation kit and observa-

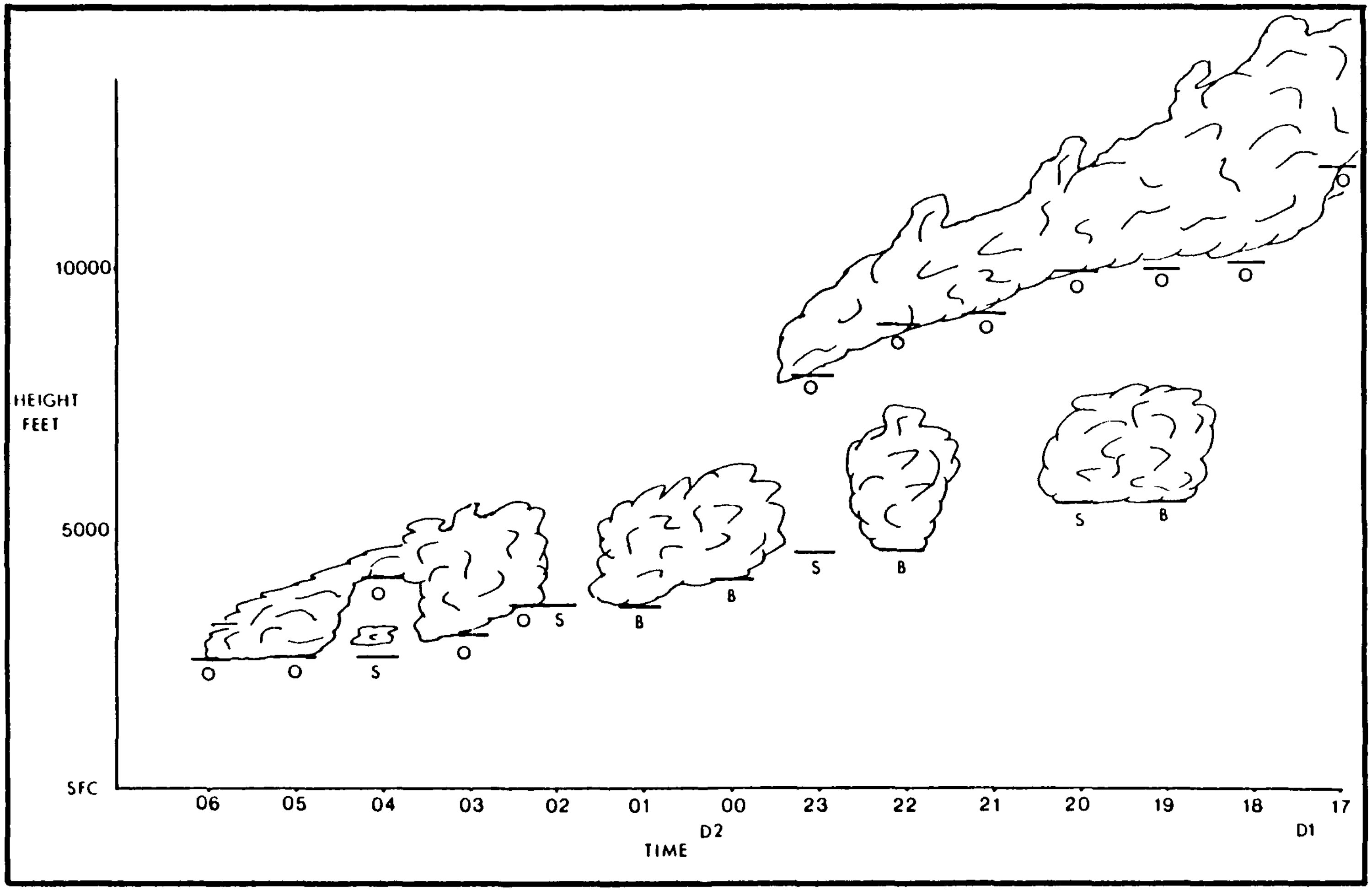


FIG. II-C-1 Time cross-section D1-1700 to D2-0600 LST. This z-t cross-section shows the cloud patterns at each hour. The — marks the base of the cloud. The O, B, and S stand for Overcast, Broken, and Scattered. The approaching warm front is indicated by the lowering of the clouds.

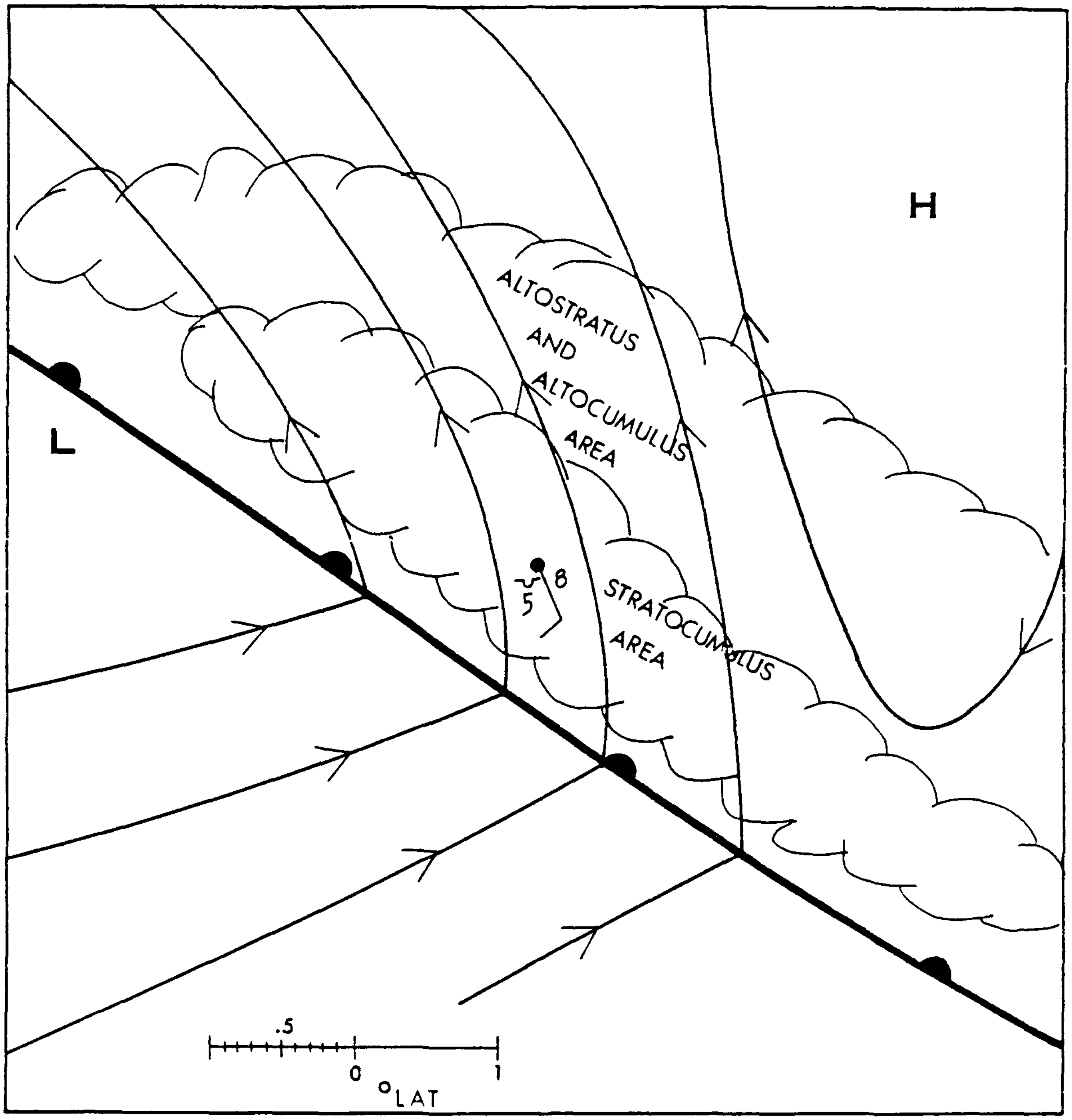


FIG. II-C-2 Surface map for D2-0600 LST. Streamlines are used to represent wind because the pressure is not known.

tions were continued using the instruments. The observations are listed in Table II-C-2. The forecaster interpreted the data as follows.

a. Sky. The clearing trend and becoming clear by D2-0856 LST indicated that the warm front was close enough that the depth of the cold air was less than the lifted condensation level.

b. Front. By D2-1259 LST the warm front was past. The wind shifted from 160° to 210°, the temperature had increased (but might be caused by the diurnal heating), however the dew point had increased rapidly during the last 5 h, which confirmed the frontal passage.

c. Ridge Aloft. By D2-1356 the ridge at 700 mb was moving past as indicated by the winds at the altocumulus level and the winds at that level should soon become southwesterly. The approaching trough would support the frontal wave with the approaching cold front. The falling surface pressure indicated an approaching low.

d. Instability. The altocumulus observed at D2-1259 LST indicated instability aloft.

e. Surface Map. Using the D2-1758 (1800) LST data a surface map was constructed to use in the weather briefing. The map is shown as Fig. II-C-3.

   1.) warm front. The warm front had passed 5 h previously and was moving about 10-12 kt, so it should be 50-60 n mi to the northeast and have the same orientation as on the D2-0600 LST map.

   2.) gradient. The surface wind would be crossing the isobars at about 30° and the gradient level wind would be at least 150% of the surface wind speed (Davenport 1961). A table for determining the spacing of isobars is given in the Appendix Part I and the spacing is 1.9° Lat.

   3.) warm sector. Ahead of the warm front the winds would continue to be from the southeast.

   4.) clouds. The low clouds remained in the cold air ahead of the warm front.

   5.) cold front. The location of the cold front is not known. However, it should be placed on the map to remind the forecaster that it is to the northwest and will pass.

TABLE II-C-2   OBSERVATIONS STARTING AT D2 0655 LST FOR STATION #3

DAY	TIME (LST)	SKY	VSBY (miles)	WW	SEA-LEVEL PRESSURE (mb)	TEMP (°F)	DEW POINT (°F)	RH	WIND DIR (°)	WIND SPEED (KT)	SKY DETAIL N_s	CC	$h_sh_sh_s$	N_s	CC	$h_sh_sh_s$	N_s	CC	$h_sh_sh_s$	REMARKS
D2	0655	E25 OVC	7+		191	19	9		160	10	10	ST	25							
D2	0755	20 SCT	7+		187	18	9		160	9	3	FS	20							
D2	0856	CLR	7+		183	22	12		160	10										
D2	0957	300 SCT	7+		176	28	17		180	8	2	CI	300							
D2	1055	300 SCT	7+		168	32	19		200	10	2	CI	300							
D2	1158	CLR	7+		143	38	22		210	11										
D2	1251	100 SCT	7+		131	40	24		210	12	2	AC	100	1	AS	100				
D2	1356	100 SCT 300 SCT	7+		110	44	24		180	15	2	AC	100	1	AS	100	2	CI	300	AC MVG E
D2	1455	300 SCT	7+		085	46	26		180	15	2	CI	300							
D2	1557	100 SCT 300 SCT	7+		071	46	26		180	14	2	AC	100	2	CI	300				
D2	1654	E100 BKN 300 BKN	7+		057	45	26		160	15	2	AC	100	4	AS	100	7	CS	300	
D2	1758	E100 BKN 300 BKN	7+		057	44	26		180	12	3	AC	100	4	AS	100	7	CS	300	AC MVG NE/CI MVG E

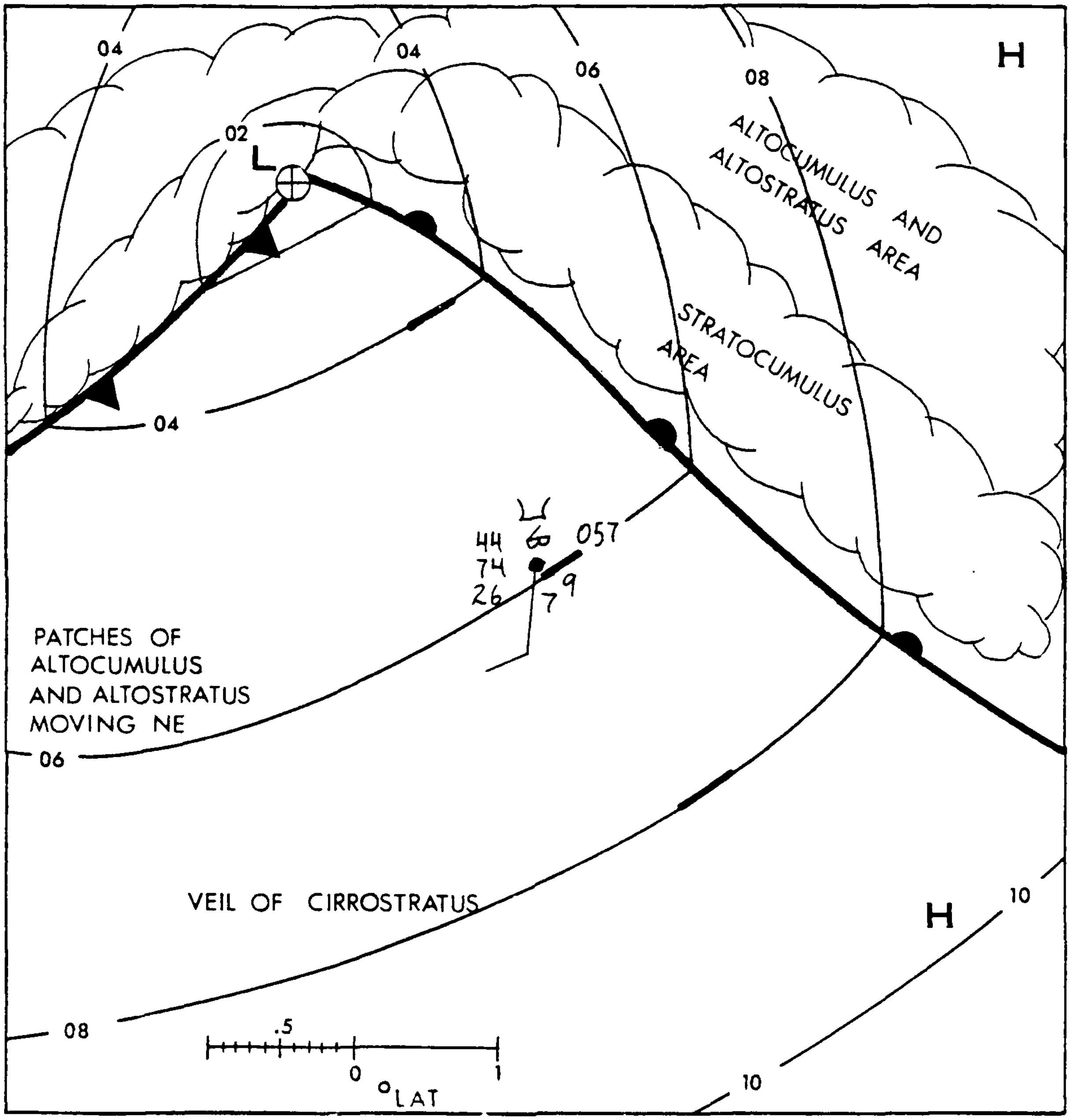


FIG. II-C-3 Surface map for D2-1800 LST. The heavy dashes on the isobars indicate the spacing of the isobars. The spacing is determined by using geostrophic winds in Appendix PART I.

6. Weather Briefing at D2-1930 LST.

   a. Warm Front. The warm front passed and weather during the day was suitable for most operations.

   b. Cold Front. A cold front is approaching. The estimated time of passage is midnight. The distance to the front is not known and the actual time of passage may be a few hours in error.

   c. Time of Front. At time of frontal passage rain showers will probably occur. (Showers often occur with cold fronts.)

   d. Clouds. Ceilings of 1000 ft. will occur with the showers and visibility will be reduced to 5 mi. (These conditions often occur with rain showers.)

   e. Winds. Gusty winds (30 kt) from the northwest should accompany the front.

   f. Temperature. Temperatures will lower to below freezing, probably 25° F. (The average minimum temperature is 24.5°F, Part H of Appendix.)

   g. Outlook. Partly cloudy and cool (mid 20's° F) during daylight hours of D3. (Continued cold air advection will offset radiational warming.)

7. Analysis of Data, D2-1800 LST to D3-0600 LST.

   a. General. The sun has set by 1800 LST but no cooling has occurred by 2200 LST. The dew point is increasing slowly and the clouds and winds are not changing. See Table II-C-3.

   b. Pressure. The pressure, however, has reached its lowest value of 1003.6 mb at 1957 LST and has started to rise. The diurnal pressure has a minimum of 1600 LST and a maximum of 2200 LST. The three-hourly tendency is shown in Table II-C-4. At 2055 LST it would appear that the decreasing pressure from the moving low is being offset by the diurnal pressure change. The warm sector isobars are still orientated from the southwest to northeast so the low is probably moving toward the northeast. That direction would correspond to the wind aloft from the D2-1750 LST observation.

TABLE II-C-3 OBSERVATIONS STARTING AT D2 1855 LST FOR STATION #3

DAY	TIME (LST)	SKY	VSBY (mi)	WW	SEA LEVEL PRESSURE (mb)	TEMP (°F)	DEW POINT (°F)	RH	WIND DIR (°)	WIND SPEED (KT)	SKY DETAIL N_1	CC	$h_1h_1h_1$	N_2	CC	$h_2h_2h_2$	N_3	CC	$h_3h_3h_3$	REMARKS
D2	1855	E120 BKN	7+		047	44	27		180	17	3	AC	100	5	AS	120				700 mb T-Td E 3°C
D2	1957	E120 OVC	7+		036	45	26		180	15	3	AC	100	7	AS	120				
D2	2055	E120 OVC	7+		046	45	27		195	14	3	AC	100	7	AS	120				
D2	2157	E120 OVC	7+		052	45	28		200	12	3	AC	100	7	AS	120				SCUD W
D2	2254	E60 BKN 120 OVC	7+		061	45	28		225	14	6	ST	60	3	AC	100	7	AS	120	LYR ST 180-040
D2	2355	E30 BKN 45 OVC	7+	RW-	064	45	30		225	12G17	6	ST	30	10	NS	45				RB 35
D3	0056	E50 OVC	7+	RW-	059	44	34		225	15	10	NS	50							NS MVG ENE 40 KT
D3	0155	E30 OVC	7+		062	44	35		270	15	10	ST	30							RE 25
D3	0215	E17 BKN 30 OVC	6	RW-					295	18G25	6	ST	17	10	NS	30				RB 13
D3	0242	E10 OVC	7+						300	15	10	ST	10							RE 42 ST MVG SE 30 KT
D3	0255	10 SCT E15 OVC	7+		077	41	39		315	13	3	ST	10	10	SC	15				RB 6 RE 25
D3	0319	E9 OVC	7+						290	20										
D3	0355	E9 OVC	7+		094	38	36		315	18G27	10	SC	9							
D3	0456	E15 OVC	7+		113	34	29		340	20	10	SC	15							
D3	0534	E13 OVC	7+						340	18	10	SC	13							
D3	0554	E13 OVC	7+		123	32	29		315	20	10	SC	13							SC MVG SE 30 KT

TABLE II-C-4

Diurnal pressure change at Station #3.

The numbers represent the net change in millibars during the previous 3 h.

Time LST	0000	0300	0600	0900	1200	1500	1800	2100
$\Delta p/3h$	-0.1	0.0	0.0	+1.0	-1.2	-0.9	+0.7	+0.5

c. Low Clouds West. The scud to the west at D2-2157 LST should be interpreted as the approach of the clouds associated with the cold front.

d. Stratus Clouds. The stratus clouds have moved in and have covered all except the southeast quadrant by the D2-2254 LST observation. Temperature and dew point holding steady. The rising pressure exceeding the diurnal change indicates that the low is not approaching the station, but is moving past to the north, and may be filling.

e. Showers. Light rain showers started at D2-2235 LST. Evaporation of the showers into the air lowered the ceiling rapidly. There must be some cumulus clouds but they are hidden by the lower stratus. No lightning or thunder occurred with any of the showers, so there was no deep convection activity. The climatic record (Appendix, Part H) shows thunderstorms are rare in February.

f. Front Approaching. The movement of the stratus toward the ENE at 40 kt is a good indicator that the front is close, but has not passed and that the rain showers are pre-frontal.

g. Wind Shift. By D3-0155 the wind had shifted to 270° and the front was starting to pass but the temperature and dew point had not changed.

h. Frontal Passage. By D3-0255 LST the front has passed. The wind shift to 315°, the lowering temperature and dew point and the rising pressure of 1.7 mb in one hour at a time the diurnal pressure is essentially steady, are enough to indicate that the front passed. The rain shower and gusty winds at D3-0215 LST marked the passage.

The low clouds, cooling temperature and finally the cloud direction at D3-0554 LST all are features of the cold air arriving.

i. The D3-0600 LST Surface Map. Construction of the D3-0600 LST map would show the following features. (See Fig. II-C-4.)

1.) cold front. The front is orientated by the shear vector (see Chapter III) using the surface wind and the motion of the alto-cumulus layer at D2-1758 LST which provides the latest estimate of wind aloft. The distance the front is past the station is determined by the wind normal to the front. The front passed 4 h ago and with a 20 kt wind (20 kt x 4 h = 80 n mi ≅ 1.3° L).

2.) isobars. The spacing of the isobars is taken from the geo-strophic wind tables and with the gusty winds up to 27 kt and the estimated 1300 ft winds at 30 kt would give a spacing of 1.2° Lat.

3.) warm front. The warm front was extrapolated from the past two maps and carried as a reminder.

4.) showers. The showers were placed along the front because show-ers were with the front as it passed.

5.) clouds. The cloud patterns also were carried for the same reason, i.e., time continuity. The edge of the stratified layer was indicated to be to the northwest by using the wave model (Fig. II-B-1b). Clearing usually occurs soon after a frontal passage.

8. Weather Briefing at D3-0730 LST. The forecast for the next 12 h is:

a. Sky Condition. By D3-0700 LST skies have cleared. However, with the front to the south, middle clouds may form above the front, and cirrus will be present. The forecast is 10,000 ft broken and 35,000 ft overcast for the next 12 h.

b. Precipitation. Good chance of snow during afternoon. No significant amounts. (Weak overrunning over the front will cause some light precipitation and with the temperature below freezing, snow.)

c. Visibility. 7+ mi next 12 h, except down to 5 mi during snow.

d. Wind. The winds with the shift from northwest to northeast and become 10-15 kt as the high pressure center moves to the north of the station.

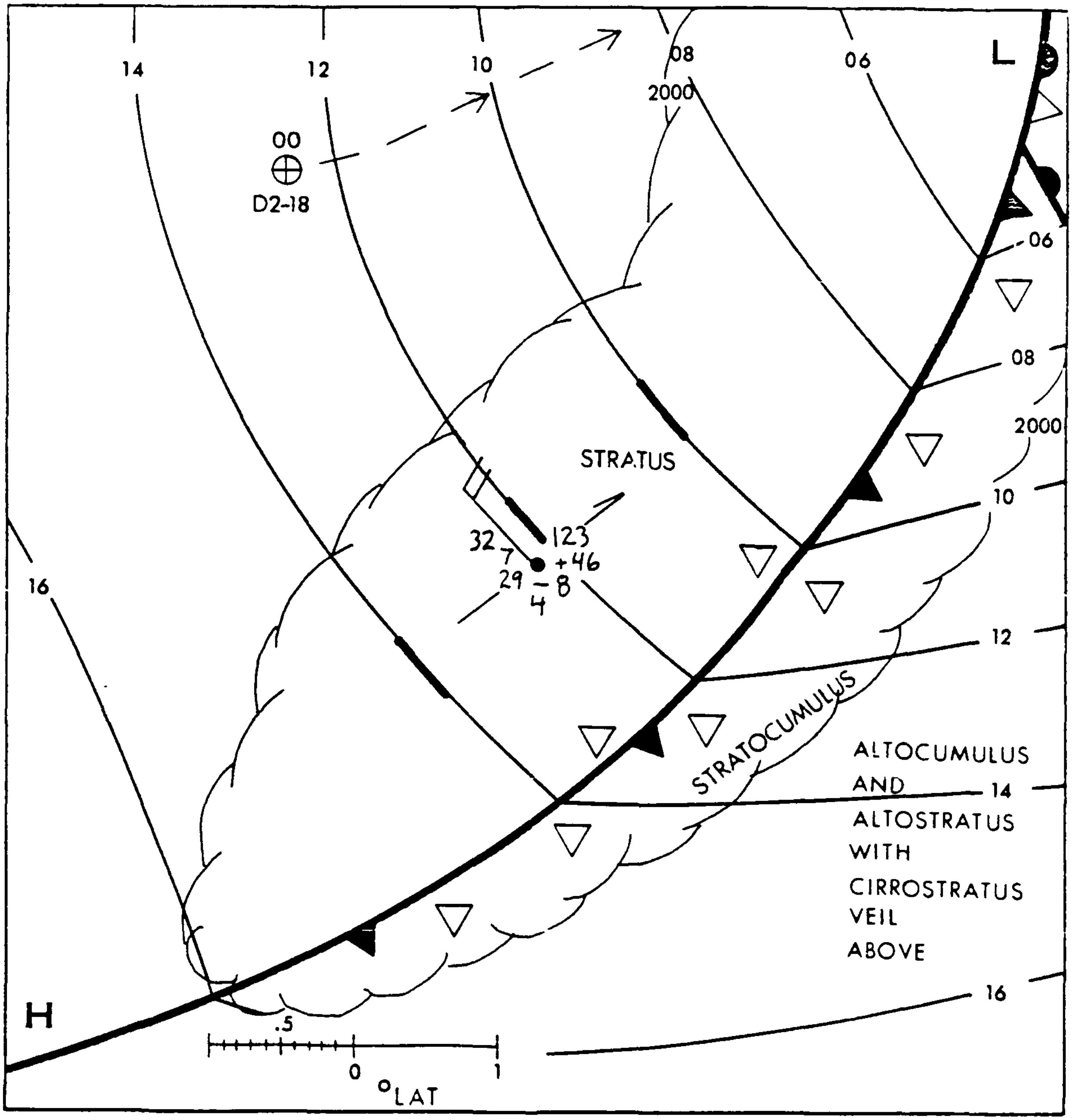


FIG. II-C-4 Surface map for D3-0600 LST. The heavy dashes on the isobars are the spacing using the wind tables in Appendix PART I. The ⟶ represents the shear vector which approximates the orientation of the front.

e. Temperature. Upper 20's during the day, equivalent chill temperature of 10° F.

f. Outlook. The front will start to return as a warm front and a frontal wave will approach from the west. Lowering of the ceilings and a good chance of snow and/or freezing rain. Temperature will remain 23-28° F. Winds becoming easterly at 10 kt.

CHAPTER III

CLASSICAL SINGLE STATION ANALYSIS AND FORECASTING

PART A THEORY

1. Introduction. In this chapter the classical single station analysis will be described. The weather station is intact with all surface observation equipment operational. The rawinsonde data are available but radar and satellite data are not. The basic procedure was described by Oliver and Oliver (1945) in the Handbook of Meteorology. However, since 1945 a change has been made from the representation of meteorological data on constant level surfaces to constant pressure surfaces. Thus, while the basic concepts involved in SSA expressed by Oliver and Oliver have not changed, the pertinent equations, maps and diagrams to be used have changed.

   This part will discuss the theory of winds of the boundary layer, gradient and geostrophic winds, and concepts related to the thermal wind.

2. Winds in the Boundary Layer. The earth's surface exerts a frictional drag on the air as it moves over the surface. As a consequence, near the earth's surface the velocity is subgeostrophic and the wind vector crosses the isobars (or height contours of isobaric surfaces) with a component from high to low values. However, the influence of the earth's surface on the flow diminishes with height so that the velocity vector approaches the geostrophic wind vector with height. The level at which the direction of the wind first becomes the same as that of the geostrophic wind is called the gradient level. An idealized (theoretical) version of the wind distribution in the boundary layer is shown in Fig. III-A-1. The arrows in this figure represent wind vectors at the ground and at 1000 ft increments above the ground. The line passing through the tips of the vectors exhibits a spiral shape which is called the Ekman Spiral. The last vector in the spiral in Fig. III-A-1 is the wind at the gradient level and is the geostrophic wind for the frictional boundary layer. This will be referred to as the gradient level wind (GLW).

   The conditions that lie behind the theory of the Ekman Spiral will generally not exist at the time that an upper air sounding is made. Thus, a plot of real winds on a hodograph will only approximate the

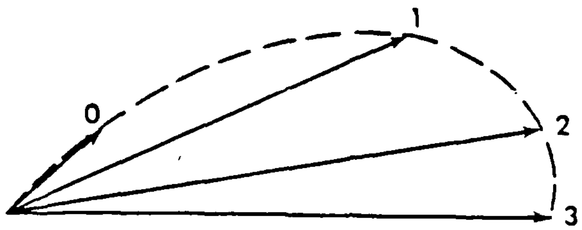


FIG. III-A-1 Hodograph for the Ekman Spiral. This illustrates how the wind vectors at successive heights approach the geostrophic wind for the boundary layer. The 3000 ft wind (3) is considered to be the GLW.

distribution shown in Fig. III-A-1. Observations show that the angle between the surface wind and the isobars on a surface map may be as small as 10° under conditions of unstable lapse rate and as large as 40° under stable conditions. The height of the gradient level averages about 1 km, but may be lower for stable conditions and higher for unstable conditions. A frontal inversion also may force an adjustment in the selection of the level of the GLW. The winds may be affected by some topographic features and this must be kept in mind when attempting to determine the GLW from the observations. As is frequently observed, the surface winds blows more nearly normal to the sea-level isobars where cold-air advection is occurring as may be seen behind cold fronts. However, the cross-isobar angle usually is diminished where warm-air advection is occurring. It would be a most unusual circumstance, perhaps only in the vicinity of a severe thunderstorm, that the surface wind would ever have a component opposite to the GLW. An informative analysis of the effects of baroclinity, stability and wind speed on the vertical wind profile in the boundary layer is given by Hoxit, 1973.

It follows from the above discussion that the wind at 3000 ft above the surface frequently should serve as a reasonable first approximation to the GLW for many circumstances. Its orientation indicates the orientation of the sea-level isobars and its magnitude gives the spacing of these isobars through the geostrophic wind relationship,

$$V_g = \frac{R_d\ T}{fp} \frac{\Delta p}{\Delta n}, \qquad \text{(III-1)}$$

where R_d is the gas constant for dry air, T is the temperature, f is the coriolis parameter, p is the pressure, Δp is the desired isobar increment, and Δn is the spacing between the isobars. Frequently, the sea-level pressure map is converted into a 1000 mb map by noting that a pressure increment of 3 mb is approximately equal to a height increment of 30 gpm. The determination of Δn may then conveniently be done by using a geostrophic wind nomogram, or it can be calculated from III-2:

$$\Delta n = \frac{0.104\ T}{V_g \sin \phi} \text{ [degrees of latitude]}, \qquad \text{(III-2)}$$

where T is temperature in °K, V_g is the observed wind at the gradient

level in kt, and ϕ is the latitude of the station. A table for Δn based on a 2 mb increment and standard sea-level values of T and p is Part I of the Appendix.

3. The Gradient Wind. The assumptions that must be applied to the equations of motion in order to derive the geostrophic wind equation are that the motion is horizontal, frictionless and unaccelerated. Thus, the parcels must flow in a straight line at constant speed and must be above the frictional boundary layer to satisfy these requirements. However, atmospheric flow is generally along curved paths, which means that centripetal acceleration must be involved, at least.

The wind that meets the conditions of frictionless, curved flow which is parallel to the height contours of the pressure surfaces is called the gradient wind, V_{gr}. Its direction at a point is the same as V_g but its magnitude depends upon the coriolis parameter, $f = 2\Omega \sin\phi$, the magnitude of the geostrophic wind, V_g, and the radius of curvature, R, of the height contours of the pressure surface at the point. The latter is used as an approximation for the radius of curvature of the parcel's trajectory which generally is difficult to determine. For everything else the same, the value of V_{gr} for a parcel following a path which is counterclockwise around the center of curvature (cyclonic curvature) differs from that of a parcel following a clockwise path around the center of curvature (anticyclonic curvature). Also, $V_{gr}<V_g$ for cyclonic curvature, but $V_{gr}>V_g$ for anticyclonic curvature. Thus, it follows that if the real wind is in gradient balance, an error will be made in determining the spacing of height contours in the vicinity of that wind by treating it as a geostrophic wind.

The equations which relate the magnitude of V_g to the magnitude of V_{gr} are as follows,

$$\text{Cyclonic Curvature:} \quad V_g = V_{gr} + \frac{V_{gr}^2}{fR}, \qquad \text{(III-3)}$$

$$\text{Anticyclonic Curvature:} \quad V_g = V_{gr} - \frac{V_{gr}^2}{fR}. \qquad \text{(III-4)}$$

The magnitude of the geostrophic wind is given by

$$V_g = \frac{g}{f}\frac{\Delta z}{\Delta n},$$

which can be solved to obtain the distance Δn between two height contours differing in value by Δz (e.g., 30 m). Thus,

$$\Delta n = \frac{g\ \Delta z}{fV_g}. \qquad \text{(III-5)}$$

Therefore, a better estimate of the spacing of height contours in curved flow can be obtained by using the observed wind, V, as V_{gr} on the right-hand side of (III-3) or (III-4) to solve for V_g. This value of V_g used in (III-5) will give the proper spacing. The results of combining (III-5) with (III-3) and (III-4) for Δz = 30 m are,

$$\text{Cyclonic Curvature:} \quad \Delta n = \frac{294R}{8.32VR\ \sin\phi + 26.46 \times 10^{-2}V^2}, \qquad \text{(III-6)}$$

$$\text{Anticyclonic Curvature:} \quad \Delta n = \frac{294R}{8.32VR\ \sin\phi - 26.46 \times 10^{-2}V^2}. \qquad \text{(III-7)}$$

In using these equations, the units of V are kt, the units of R are degrees of latitude and it is a positive number, and the units of Δn are degrees of latitude. R can be measured on the map, given the general shape of the contour which passes through the observation point. Thus, a calculation of Δn can be made. For an example calculation, suppose that the observed wind is 50 kt, the latitude is 45° and the radius of curvature is 9° Lat. Treating the wind as geostrophic, the value of Δn calculated from (III-5) is 1° Lat (see Appendix, Part J). But, from (III-6), the calculated value of Δn is 0.8° Lat and from (III-7), it is 1.3° Lat.

Obviously, when doing single station analysis, a value for R cannot be obtained; however, it may be possible to determine whether the curvature is cyclonic or anticyclonic. If so, the sample calculations provide a guideline for modifying the spacing obtained from (III-5) or Appendix, Part J. The rule is to make the spacing a little larger if the curvature is anticyclonic and a little smaller for cyclonic curvature.

A nomogram devised by J. C. Bellamy for calculating V_{gr} is shown in Fig. III-A-2 and in the Appendix (Part K). It also can be used to make the calculations of Δn above by following the arrows shown in the

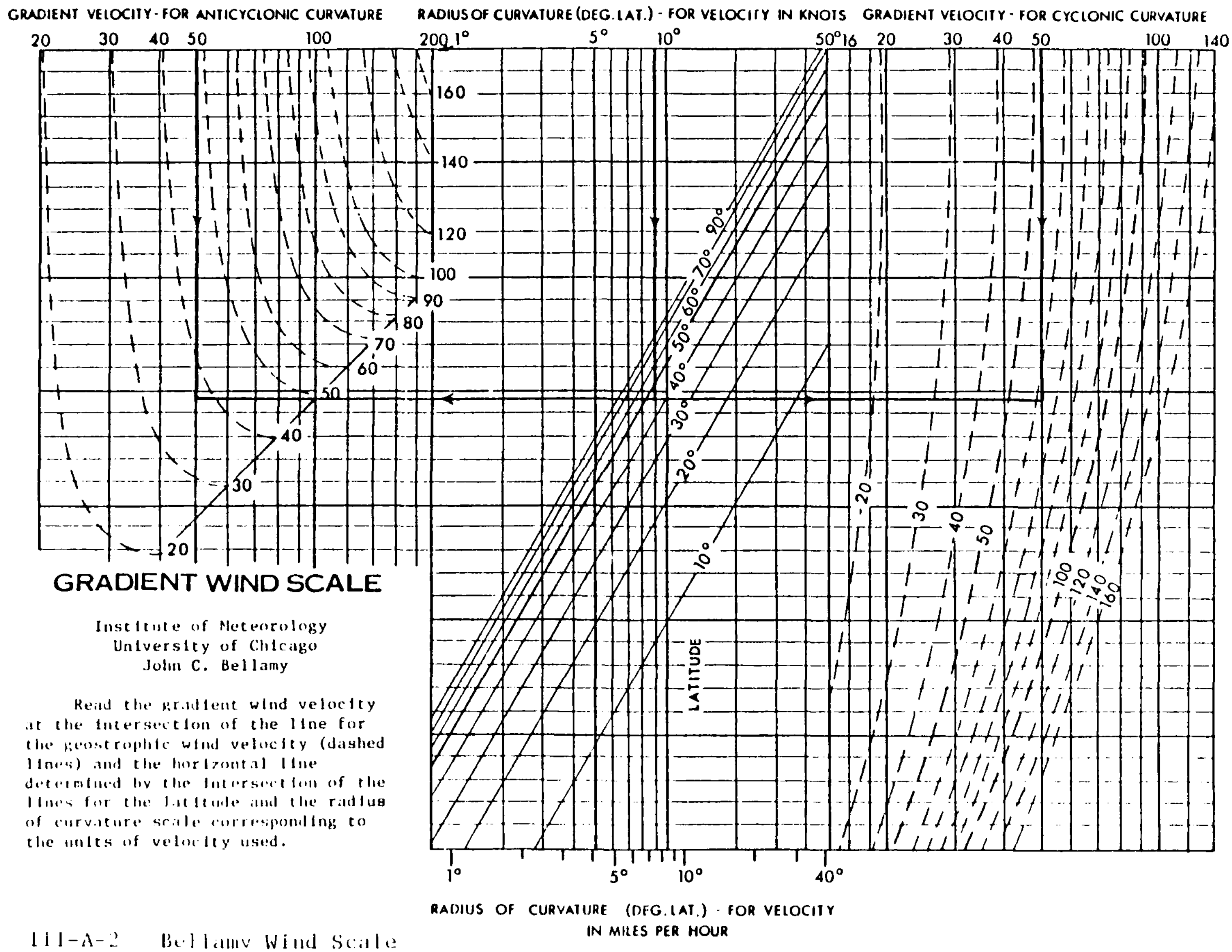


FIG. III-A-2 Bellamy Wind Scale

diagram. The nomogram (Fig. III-A-2) is entered in the middle section with R = 9° Lat. The vertical arrow locates the intersection of the lines for R = 9° Lat and ϕ = 45°. Moving from this point to the right in the diagram to the intersection with the vertical line labeled as 50 kt then determines V_g = 63 kt (dashed lines) for cyclonic curvature. With V_g = 63 kt and ϕ = 45°, enter Appendix, Part J, and read Δn = 0.8° Lat. For anticylconic curvature of R = 9° Lat, move to the left in the nomogram to the intersection with vertical line for V_{gr} = V = 50 kt. The dashed lines can be interpolated to give V_g = 38 kt at this point. Then, enter Appendix, Part J, with this value and ϕ = 45° to read Δn = 1.3° Lat.

4. <u>The Thermal Wind</u>. Physically, the concept of a thermal wind arises from the combination of the influence of the temperature field on the pressure field, as given by the hypsometric equation, and the control exerted by the pressure field on the wind field, as given by the geostrophic wind equation. The hypsometric equation

$$\Delta z = \frac{R_d \overline{T}_v}{g} \ln \frac{p_1}{p_2} , \qquad \text{(III-8)}$$

states that the vertical distance in going from pressure surface p_1 up to pressure surface p_2 is proportional to the mean virtual temperature $\overline{T}_v$ in the layer. Thus, where the air in the layer is warm, Δz will be relatively large and where $\overline{T}_v$ is small the thickness will be small. This, in turn, means that the slope of the pressure surfaces must change with height over most areas of the middle latitudes, mainly because of the horizontal variation of temperature.

The vector form of the geostrophic wind equation for use on isobaric surfaces is given by

$$\vec{V}_g = \frac{g}{f} \hat{k} \times \nabla z , \qquad \text{(III-9)}$$

which says that $\vec{V}_g$ at a point is parallel to the height contours with high contour values to its right and its magnitude is proportional to the slope of the given pressure surface at the point. Thus, where the slopes of the pressure surfaces are changing in magnitude and orientation with height, the geostrophic wind also is changing in magnitude and orientation with height. Since above the friction layer the geostrophic wind is a first approximation to the real wind, it follows that the real wind changes with height in essentially the same way as $\vec{V}_g$.

The vector representing the shear of the geostrophic wind on going from surface p_1 to surface p_2 is called the thermal wind and is given by

$$\vec{V}_t = \frac{g}{f} \hat{k} \times \nabla(\Delta z) \quad . \tag{III-10}$$

This vector is oriented parallel to the thickness lines for the layer from p_1 to p_2 with higher value of thickness to its right--therefore from (III-8) $\vec{V}_t$ has higher values of $\overline{T}_v$ to its right. From this, when the wind veers with height, there is warm air advection in the layer and cold air advection is occurring in a layer in which the wind backs with height. These concepts are applicable only at levels above the gradient level.

From the combination of (III-8) and (III-10), one can develop an equation to provide the spacing of $\overline{T}_v$ on a pressure surface, based upon the thermal wind in a layer which encompasses the given pressure surface. If the thickness of the layer is kept to a few thousand feet, this spacing will be representative of the actual isotherms. The resulting relationship is

$$\Delta n = R_d \ln(\frac{p_1}{p_2}) \frac{\Delta\overline{T}_v}{f \; V_t} \quad .$$

Another formulation of this equation is given by

$$\Delta n = \frac{g \; \Delta z \; \Delta\overline{T}_v}{f \quad V_t \; \overline{T}_v} \quad . \tag{III-11}$$

Here, Δz is the vertical distance over which the shear vector is determined. Part L in the Appendix gives values for Δn for several choices of $\Delta\overline{T}$ and heights. For this purpose, the correction of the temperature to the virtual temperature, T_v, is unimportant and standard atmosphere values for $\overline{T}$ have been used.

In Fig. III-A-3, suppose that the vectors represent the observed wind at 3000 ft, 6000 ft and 10,000 ft as they would appear on a hodograph. Arrows are not shown for the intermediate heights. The dashed wind shear vectors then represent an approximation to the thermal wind in the layers 3000-6000 ft and 6000-10,000 ft. In the lower layer the wind is veering and warm air advection is occurring; the backing wind in the upper layer indicates that cold air advection is present. The effect of

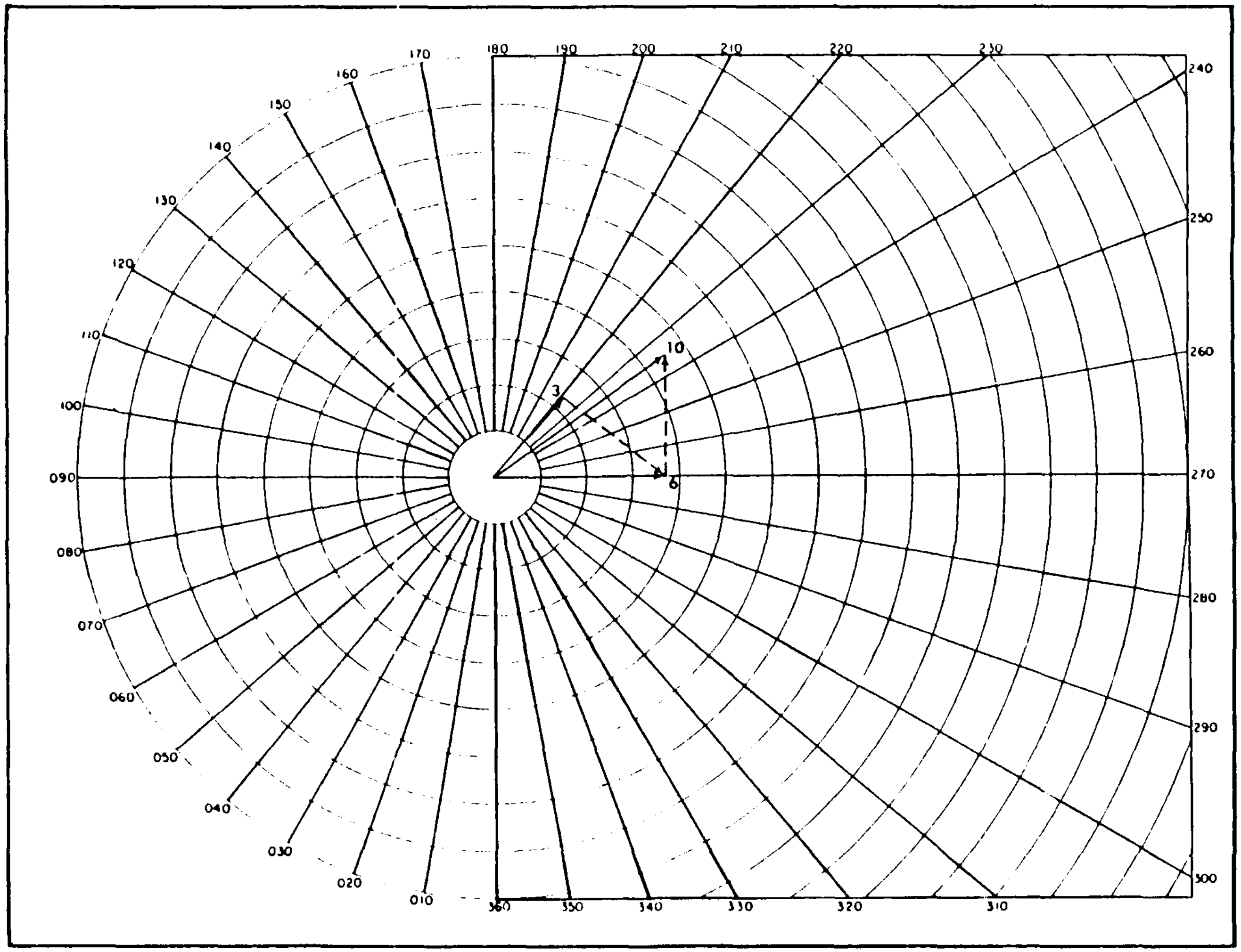


FIG. III-A-3 Hodograph for winds at selected levels. Winds at 3000 ft (assumed gradient level), 6000 ft and 10,000 ft are solid vectors. Dashed vectors represent the shear vector for the layers 3000-6000 ft and 6000-10,000 ft.

the differential advection on the lapse rate is to cause the stability to decrease with time.

The relative orientations of the shear vectors provide additional information on current conditions in the vicinity of this station. In the lower layer warm air lies off to the southwest of the station and colder air lies off to the northeast. In the upper layer, the warm air lies to the east of the station while colder air is to the west. The colder air overlying the warm air to the west of the station should be the region where the lapse rate indicates the least stability and the most likely area of convective activity. The region most likely to have stable lapse rates then lies to the east of the station where, relatively speaking, warm air overlies cold air.

In Figs. III-A-4a and III-A-4b are shown the wind vectors $\vec{V}_1$ and $\vec{V}_2$ for the bottom and top pressure surfaces, respectively, of a given layer. The shear vector, i.e., the thermal wind, is labeled as $\vec{V}_t$. The vector labeled $\vec{V}_n$ is the component of the wind in the layer which provides the temperature or thickness advection. In Fig. III-A-4a, $\vec{V}_n$ is normal to $\vec{V}_t$ and in the same direction as the gradient of $\overline{T}_v$, i.e., toward the warmer air. This is considered to be the positive direction for $\vec{V}_n$. If it is assumed that the local change of thickness is a consequence only of horizontal advection, it can be shown that

$$\frac{\partial \Delta z}{\partial t} = -\frac{f}{g} V_n V_t \quad . \qquad \text{(III-12)}$$

In this example, cold air advection is occurring and the thickness is decreasing with time, as indicated by the minus sign in the equation.

The wind is seen to be veering in Fig. III-A-4b and $\vec{V}_n$ is in the opposite direction to the gradient of mean virtual temperature. This makes the magnitude of $\vec{V}_n$ a negative quantity in the thickness advection equation which, therefore, gives a computation of a positive thickness change.

In practice, (III-12) can be used to tie a forecast of height fields at two pressure levels together in a hydrostatically consistent fashion. For example, if a forecast is made of the 1000 mb surface, (III-12) could be used to determine the new thickness from 1000 to 500 mb at a number of locations. However, because the equation does not account for vertical

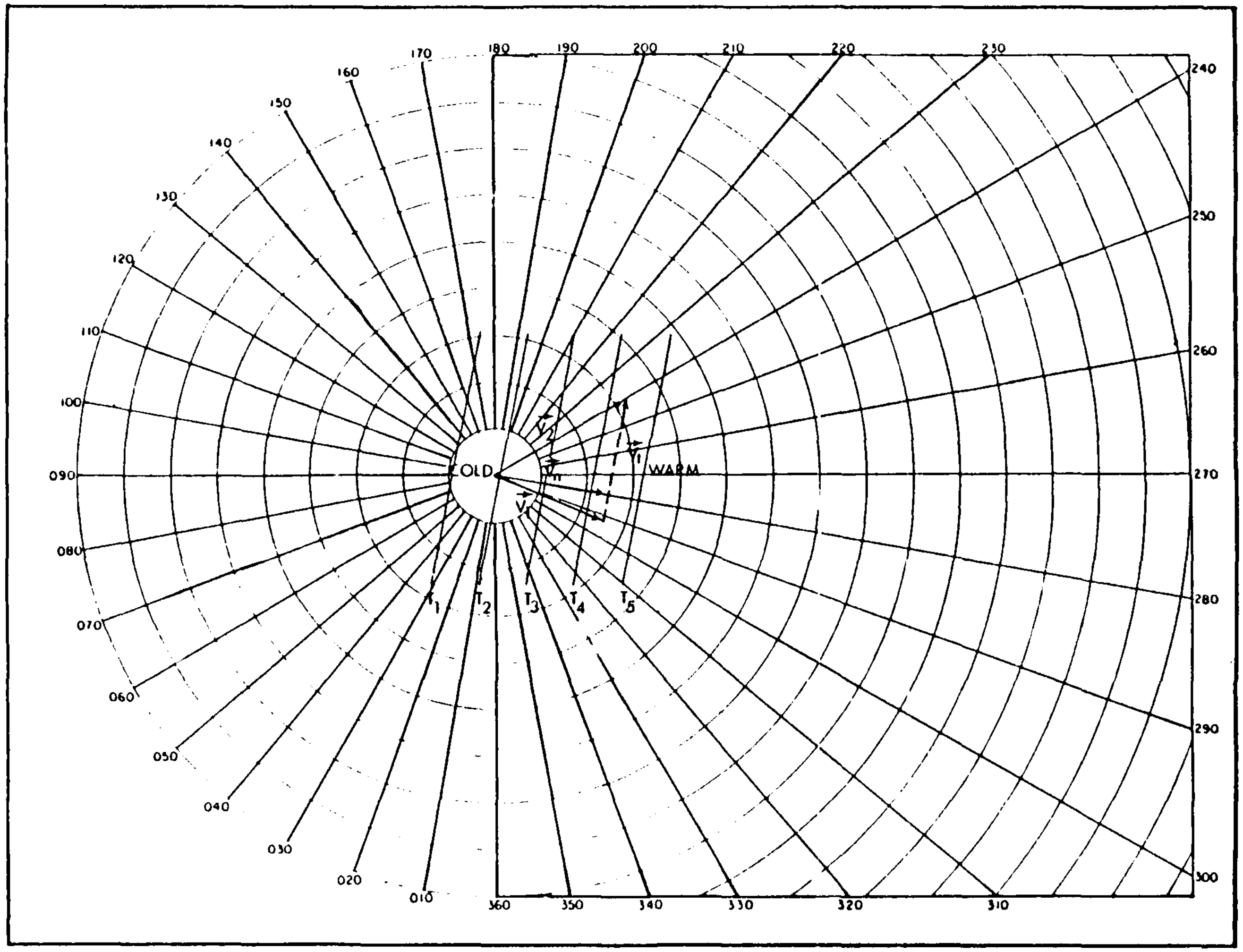


FIG. III-A-4a Hodograph for backing wind. The vectors $\vec{V}_1$ and $\vec{V}_2$ show the wind to be backing from the bottom to the top, respectively, of a given layer. The orientation of the shear vector (dashed) indicates the relative locations of warm and cold air. $\vec{V}_n$ is the wind component in the layer which provides cold air advection.

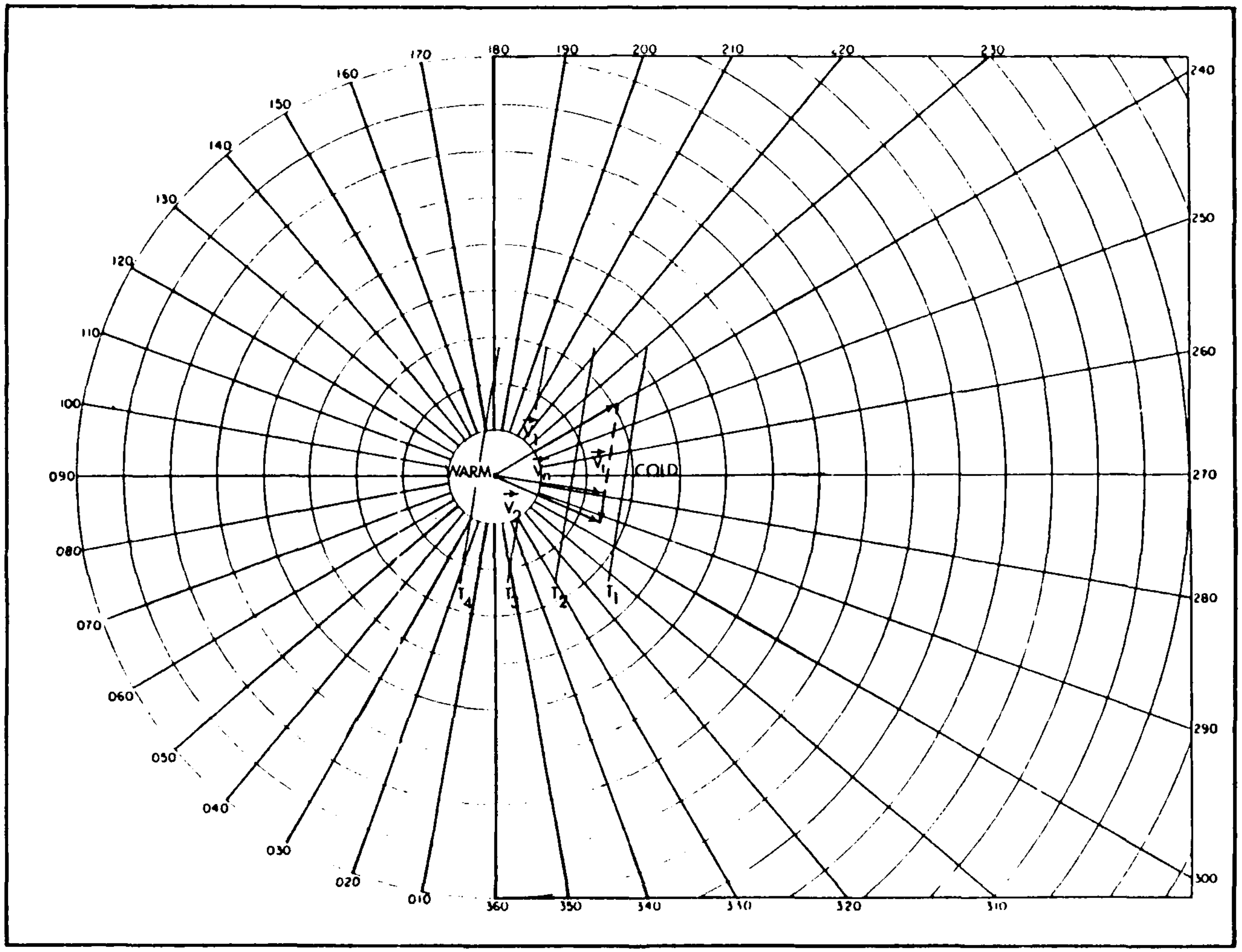


FIG. III-A-4b Hodograph for veering wind. The vectors $\vec{V}_1$ and $\vec{V}_2$ in the layer show a veering wind and, thus, warm air advection.

advection or adiabatic changes of state associated with vertical displacement and non-adiabatic processes, it tends to predict thickness change which is too large by a factor of about two. If a factor of two correction is made in (III-12), a predicted value of thickness in gpm is given by

$$\Delta z = \Delta z_o - 4.8 \times 10^{-3} V_n V_t \Delta t \qquad \text{(III-13)}$$

The subscript zero indicates the value of thickness at the initial time, Δt represents the number of hours of the forecast and the units of V_n and V_t are kt.

Warm air advection near the ground may be an indication of an approaching warm front, especially if it extends over a deep layer. Cold air advection may indicate that a cold front already has passed and is moving away. Warm air advection also precedes occluded fronts and, although of weaker magnitude, may occur in the warm sector of a warm cyclone. These alterations in advection, coupled with the information provided by the surface observations, will enable the single-station analyst to estimate his position relative to fronts. If he finds that an approaching warm front has small advection in the 3000-6000 ft layer but large warm air advection above, he may assume the front will be active. An active cold front is indicated when the winds above the frontal boundary are either parallel to the front or blowing with a component from warm to cold.

Information on the identification of fronts and other stable layers in RAOBS in terms of their temperature, humidity and wind characteristics can be found in PART B of this chapter, Chapter 6 of AWSTR 79/006, and on pages 8-1.6-2 to 8-1.6-4 in AWSP 105-56, which are copied in the Appendix as Part M.

PART B PRACTICAL PROCEDURES AND METHODS

1. Introduction. Single station analysis requires that the analyst look at the data from a different perspective and adapt procedures to extract the maximum information from the limited data. In this section the weather station is considered to be operational but is not receiving data from other stations. The only data available are from observations at the station which are surface and rawinsonde. Since most USAF/AWS weather

stations do not take radiosonde or rawinsonde data, alternate supply of this data should be preplanned. In addition to the National Weather Service, the USA Artillery makes rawinsonde soundings. Means of transmittal of the data from the observer to the station should be established prior to the emergency. USAF/AWS mobile rawinsonde stations are available and some new lightweight radiosonde equipment can be carried in a small package, for use in the field, or as backup in the station. The materials and equipment to do SSA should be in place. A list of recommended materials and equipment is given in the Appendix, Part A.

2. Continuity. In standard weather map analysis continuity is very important. Usually, when doing SSA, a time cross-section is used to record time continuity of the station. A sample plotted cross-section is shown in Fig. III-B-1. The ordinate is pressure which can be labeled in pressure altitude if desired. Time is the abscissa and when the station is in the Northern Hemisphere progresses toward the left, and toward the right if the station is in the Southern Hemisphere. These directions are used so that the wind shifts will appear as on a surface map. The data used in the cross-section in Fig. III-B-1 starts at time D1-0600 LST which is 12 h prior to the data used in Chapter II, PART C, which started at D1-1755 LST.

   a. Data. Surface data may be plotted along the bottom of the form using standard plotting models. The winds aloft may be plotted at the dots shown on the vertical wind staffs as is done on the Skew T-Log P diagrams. The radiosonde data may be plotted with the temperature and dewpoint spread at the pressure levels and at all significant levels if desired. When the sounding also is plotted on the Skew T-Log P diagram, then the plotting of significant points on the cross-section is not necessary. Heights may be plotted at mandatory levels (see Fig. III-B-1). The cross-sections may make a continuous record. As new data are received they are added to the cross-section and analyzed.

   b. Cross-section Analysis. A number of features should be analyzed. It may be necessary to have more than one cross-section so that the analysis does not become cluttered.

      1.) advection. Advection can be shown by making red (warm) and blue (cold) marks between winds on the wind staff. No or

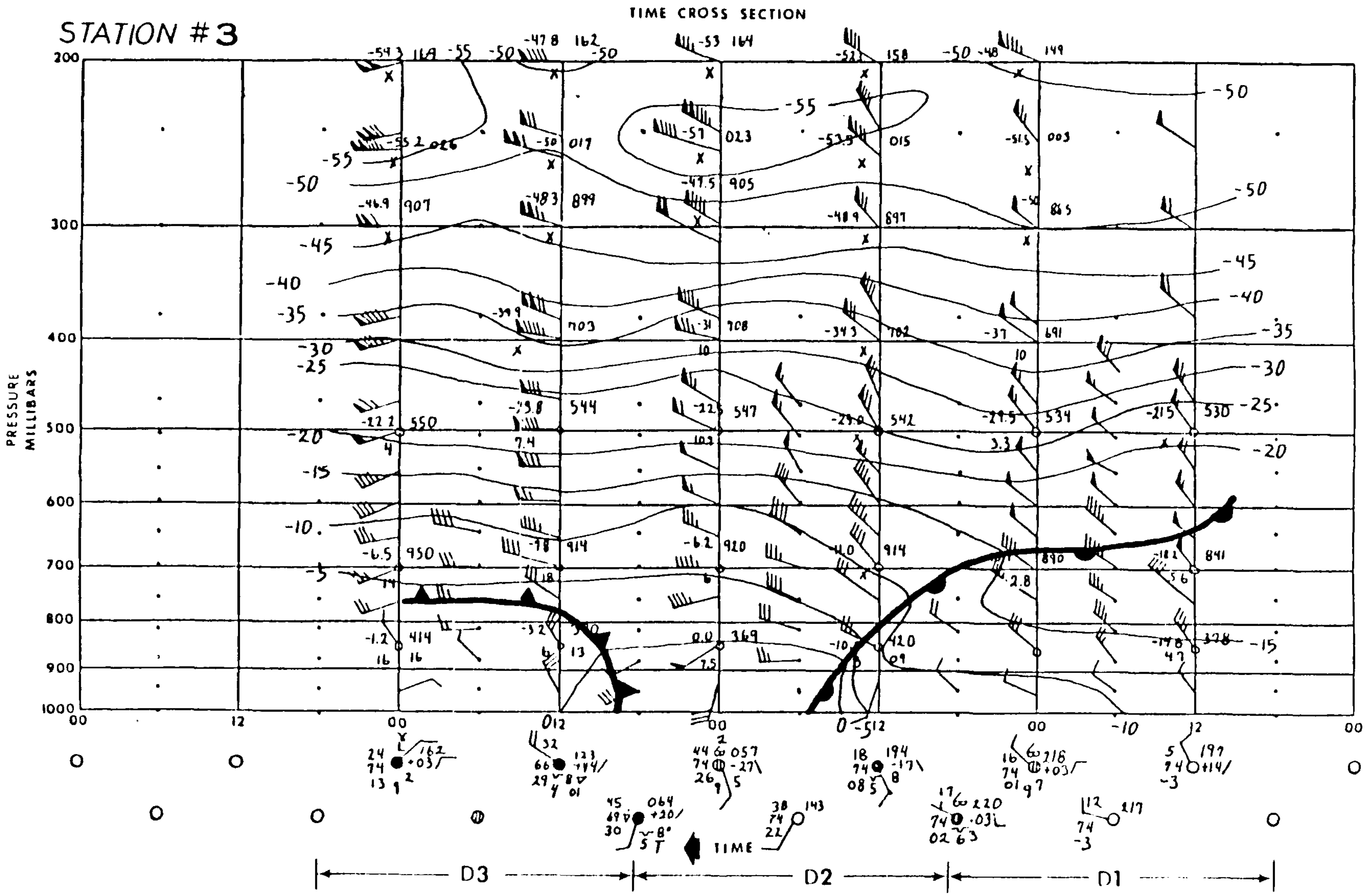


FIG. III-B-1 Time-height continuity chart showing fronts and isotherms ° C.

little advection can be left blank or marked in yellow to indicate that the analysis was made.

2.) temperature. Isotherms may be drawn on the chart, usually in red at 5° C intervals. The isotherms will show inversions by being closed at some level above the surface, will tend to become vertical at fronts, and will be nearly horizontal in air masses (see Fig. III-B-1).

3.) temperature changes $\Delta T/12$ h and $\Delta T/24$ h. A 12 and 24 h temperature change can be calculated, especially at the mandatory levels. These values are usually plotted on the cross-section close to the last observation and between the two times used (not shown). The 12 h change may include a diurnal effect, especially in the lower levels.

4.) potential temperature. The potential temperature may be computed for each pressure with plotted data. Usually the potential temperature values are read from the Skew T-Log P Diagram. Since the potential temperature is a semi-conservative value it can be very useful in both frontal analysis and in stability considerations.

5.) dew point spread. The difference between the temperature and dew point relates to the relative humidity. The RH for some $T-T_d$ values at different pressures are shown in the Appendix, Part C. Using the surface reports and the RH (dewpoint spread) aloft a cloud chart may be developed as shown in Fig. III-B-2. First all plotted raob data with $T-T_d < 4°$ C are located. Then the clouds from the surface reports were plotted at the height reported. The rain shower was the basis of the cumulus cloud drawn at D3-0000 LST. RH becomes important in relation to clouds and aircraft icing.

6.) height change (pressure change). $\Delta H/12$ h, $\Delta H/24$ h. The height change at the mandatory levels can be computed and plotted on the cross-section. The surface pressure should be converted to a 1000 mb height. One easy method is to use the nomogram on the Skew T-Log P Diagram USAF/AWS WPC 9-16 and described in AWSTR 79/006. An approximation of the change from surface pressure to a 1000 mb height is given following equation

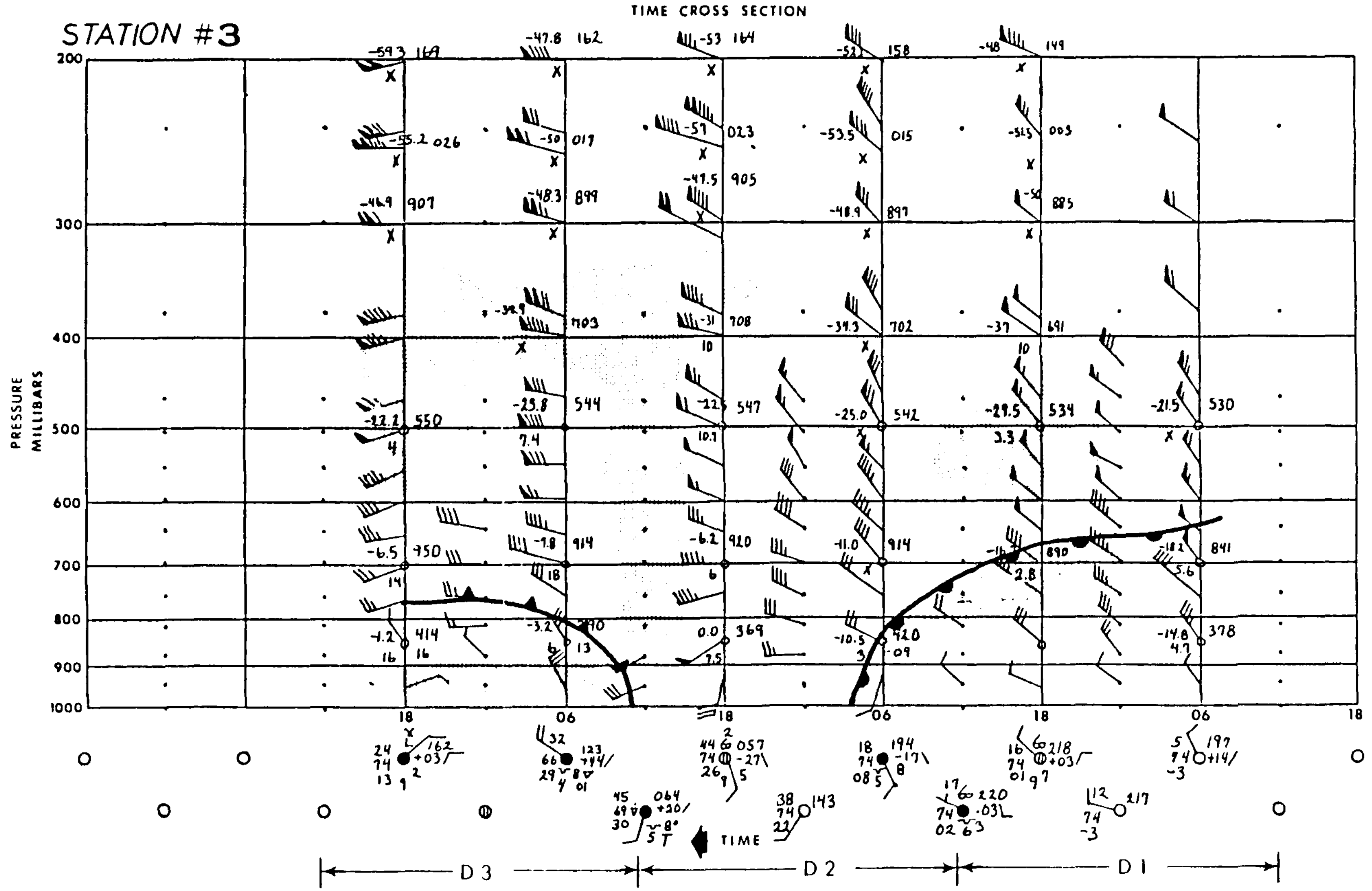


FIG. III-B-2 Time-height continuity chart showing fronts and clouds (stipled areas).

(III-1). After the surface pressure is converted to height the ΔH in meters can be computed. Such an analysis for 12-h ΔH is shown in Fig. III-B-3. The height change may indicate warm or cold advection and/or movement of systems.

7.) synoptic feature analysis.

a.) Troughs and ridges aloft will be indicated by wind shifts and the ΔH patterns. The troughs are marked in brown dashed lines and the ridges in sawtooth yellow lines. The troughs and ridges are shown in Fig. III-B-3.

b.) Fronts are indicated by red, blue and purple. They are located by using the advection, wind shifts, ΔT, ΔH (Δp), and weather--the same as with any weather map analysis. The slope of the front on the cross-section is a rate of height change, $\Delta z/\Delta t$; if calculated it would be in ft(m)/h.

8.) other features. The tropopause height may be indicated. The location of the tropopause is apparent but not marked on Fig. III-B-1. Stability indexes such as TTI, SI, KI and SWEAT may be computed and listed at the top where space is available. Other items can be added to meet operational requirements like vertical wind shear in kt/1000 ft for CAT forecasting (AWS 1979, AWSP 105-56), B values for the Refraction Index (Moreland, 1965), and the surface weather along the bottom by using conventional markings.

3. Map Analysis. To be useful in the conventional sense, the data on the z-t section must be converted to base maps which show the surface and required upper level charts. The transformation from the z-t to the x-y surface requires the data to be utilized in ways that usually are not used in map analysis, but would improve most maps if they were. Select a base map which satisfies the operational requirements with the station located approximately at the center of the map. The smallest map should be about 8° by 8° Lat and a larger map is better. Plot the data using standard plotting models on the map at the station for the surface and desired upper levels.

a. Surface Pressure Field. The isobars are drawn parallel to the gradient-level wind (GLW). As explained in PART A of this chapter,

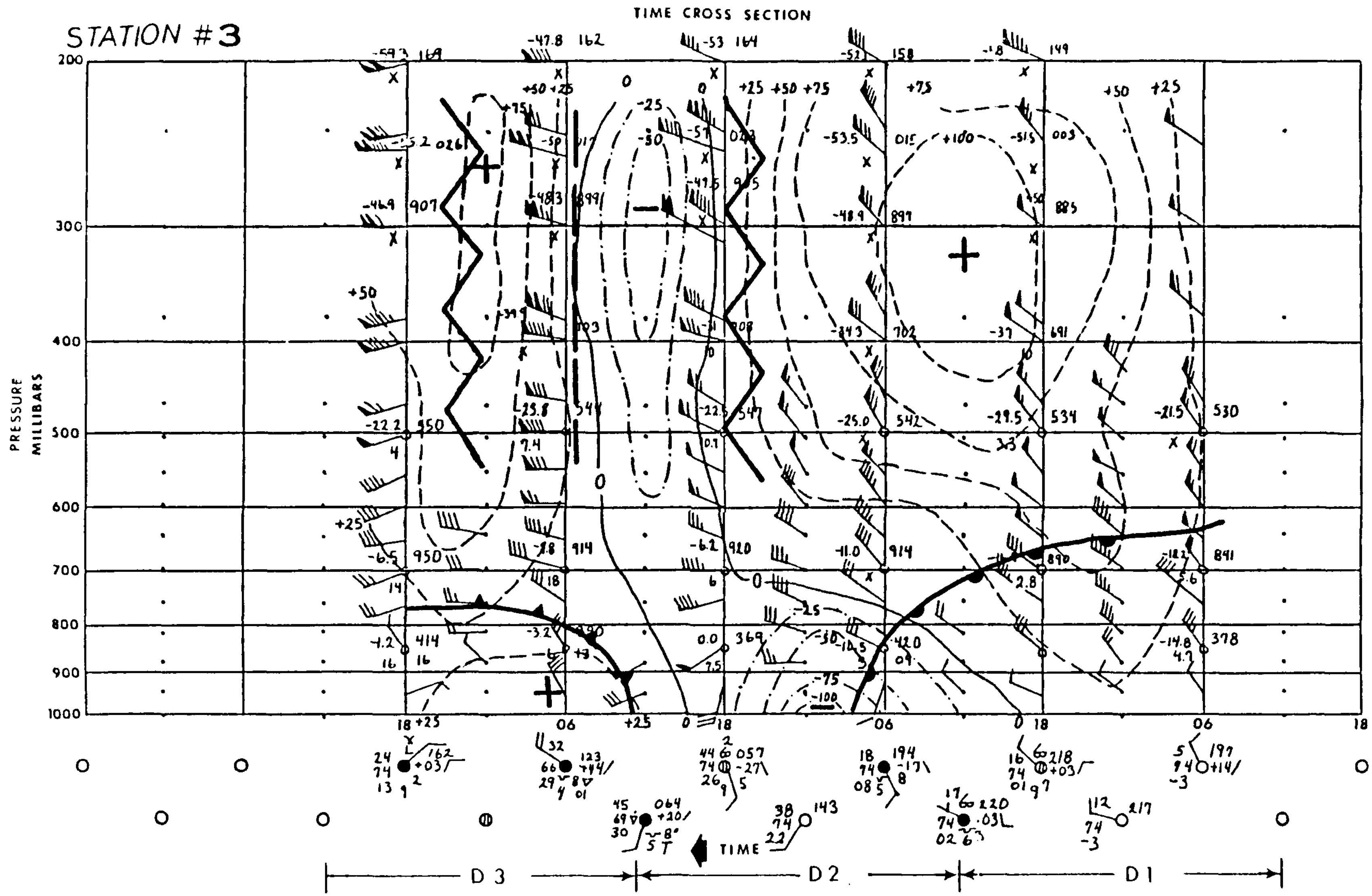


FIG. III-B-3 Time-height continuity chart showing fronts, ridges and troughs, and 12 h isallobars in meters.

the surface wind does not represent the GLW in either speed or direction. Use a wind 600 to 900 m above the surface depending upon the roughness of the surface and the stability of the atmosphere as shown by the RAOB. Plot the GLW on the surface map along with the other data. Using the sea level pressure and the GLW, three or four isobars may be drawn with the spacing determined by using a geostrophic wind scale or the table in the Appendix, Part I. Make these isobars about 2 cm long and label them in millibars. The direction to the highs and lows are now known.

Consider the pressure fields in continuity. If a low or high pressure center passed the station within the last 24 h, it is still in the area and probably has about the same central pressure. Consider how the pressure center passed the station. Say the wind was south, went to east and is now from the north, then the center of a low moving eastward went south of the station. This wind shift pattern could accompany a westward moving high pressure center which is north of the station. If the station is in the mid-latitudes this assumption is poor because not many highs move westward. More isobars may be drawn to match the known pressure field and to complete the final sketch of the map.

b. Upper Air Contour Field. The contours are parallel to the upper air winds so space the contours (Part J of Appendix) for the wind speed and draw two or three short contours in the vicinity of the station. Study the cross-section and see if the station is in a trough, ridge, or in between. Continue to draw and adjust the spacing of the contour field. Curvature will have an effect on the spacing of the contours. If the curvature is cyclonic the contour spacing will be closer than the spacing given by the geostrophic wind scale, and if the curvature is anticyclonic, the spacing must be increased. (See discussion in PART A of this chapter.)

c. Upper Air Temperature Field. The orientation of the isotherm field must be determined before the isotherms can be drawn. The isotherms are parallel to the thermal wind, but since the thermal wind is not available the vertical shear vector is used as an approximation of the thermal wind. The shear vector can be computed graphically using a hodograph or by using trigonometric functions which can be

previously programmed into a calculator. The hodograph will be used in this report because of the visual aspects. A computation of a shear vector is shown in Fig. III-A-3. The colder air is to the left of the shear vector. A selection of heights whose winds may be used to determine the shear vector for each standard level are shown in Table III-B-1.

TABLE III-B-1

Heights above MSL at which the winds may be used for computing shear vectors for standard upper-air maps. In mountainous areas the shear for the 850 mb level may not be available.

	UNITS (meters)	UNITS (feet)
Map Level	Lower/Upper	Lower/Upper
850 mb	1200-2000	4000-6000
700 mb	2400-4000	8000-12000
500 mb	4800-6000	16000-20000
300 mb	8500-11700	25000-35000
200 mb	11700-15000	35000-45000
100 mb	17000-20000	50000-60000

Tables of spacing of isotherms at the first three levels of Table III-B-1 are given in the Appendix, Part L. The spacing of isotherms at other levels may be computed using equation (III-11).

4. Air Mass Analysis. In most of North America four air masses exist with seasonal variations as to extent and characteristics. In other parts of the world similar air masses exist and have seasonal variations. The forecaster must be aware of the climatology of the area when he is doing SSA. Each air mass has a range of values of potential temperature which apply to the given geographic region. However, the ranges of values vary in relation to season of the year and are modified as the air mass moves farther from its source region. A short discussion of the more usual air masses follows.

a. cT. Continental tropical air is on the surface in the U.S. at the higher elevations and more frequently in the warmer seasons. It is found also above an inversion that may extend very far east from the plateau regions. Sometimes it is called Superior air. On the

surface the T_d is usually less than 5° C (41° F). Above the surface inversion the lapse rate aproaches the dry adiabatic lapse rate with a potential temperature of approximately 315° K (42° C) (108° F). Several soundings showing the cT air aloft are shown by Miller (1972). The boundary of cT and the cooler mT on the surface has many names (in Texas: Marfa Front, Dew Point Front, Dew Line, Dry Line) and needs to be identified on the map. The cT air mass is identified on the soundings of Figs. III-B-4 and III-B-6.

b. mT. Maritime tropical comes from the warm oceans. The mT air is important because it has the moisture for clouds and precipitation. Almost all large precipitation events involve mT air. The surface potential temperature of mT air is 300° K (27° C) (81° F). The mT air mass is shown in Figs. III-B-4 from the surface to about 800 mb and III-B-8 from the surface to 700 mb.

c. mP. Maritime polar air comes from the mid-latitude oceans and has a surface potential temperature of approximately 285° K (12° C) (54° F). A sounding close to the coast will follow a moist adiabat at least to the 500 mb level. When the mP air crosses the Rocky Mountain Range, it may lose most of its moisture, especially in the lower 10,000 ft. Fig. III-B-5 is an excellent example of the deep cooling caused by the passage of an mP front. In this example, moisture has remained in the lower levels.

d. cP and cA. Both the continental polar and continental arctic air masses are colder than the 285° K of the mP air in the winter (the summer cP is warmer than summer mP). In its source region cA air may have a potential temperature of 235° K (-40° C) (-40° F). The arctic air will modify as it moves southward. Figure III-B-6 shows both cP and cA air masses in the lower levels. The air between 680 and and 610 mb has been warmed by subsidence and has the characteristics of cT air. The cP and cA air masses are shallow (3000 m deep) while the mP may extend to the tropopause. The cooling when the mP front comes by is aloft. The cP, cA front will have cooling in the lower layers with little change aloft. On many occasions the mP and cP front (Fig. III-B-7) will be close to each other at the surface so the whole RAOB will show cooling. This is the case in Fig. III-B-7. The cP air extends up to 700 mb and the mP is aloft.

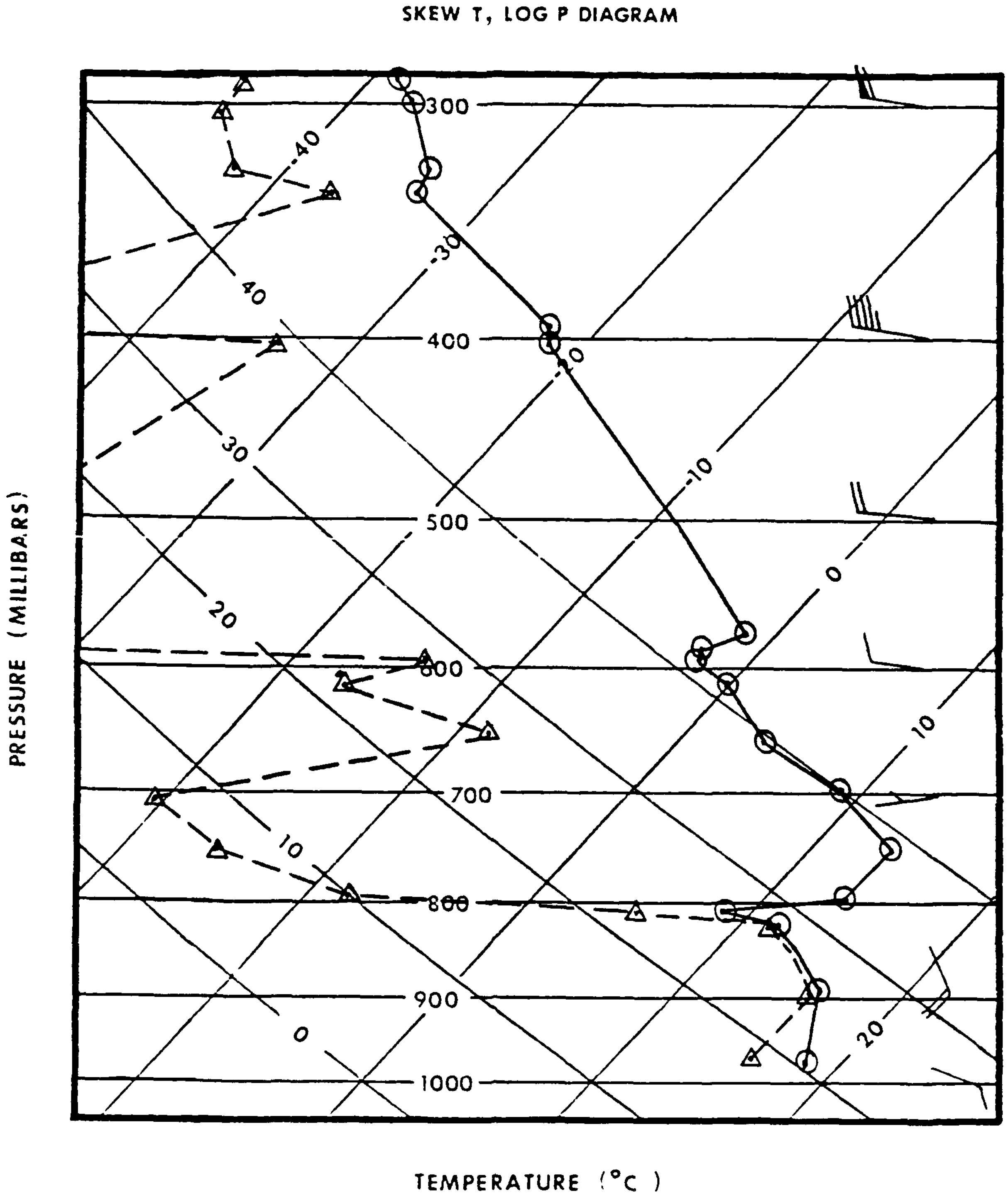


FIG. III-B-4 Del Rio, TX 72261 DRT for 1200 GMT 01 November 1983. The sounding shows the cT air above the mT air. The cT air has a potential temperature of 40° C at 700 mb and higher values above 600 mb which is caused by subsidence. The mT air between 900 and 800 mb has a potential temperature of about 26° C, and is moist. Note the super-adiabatic layer just below 800 mb. This is caused by the wet thermister approaching the wet bulb temperature rather than the air temperature.

SKEW T, LOG P DIAGRAM

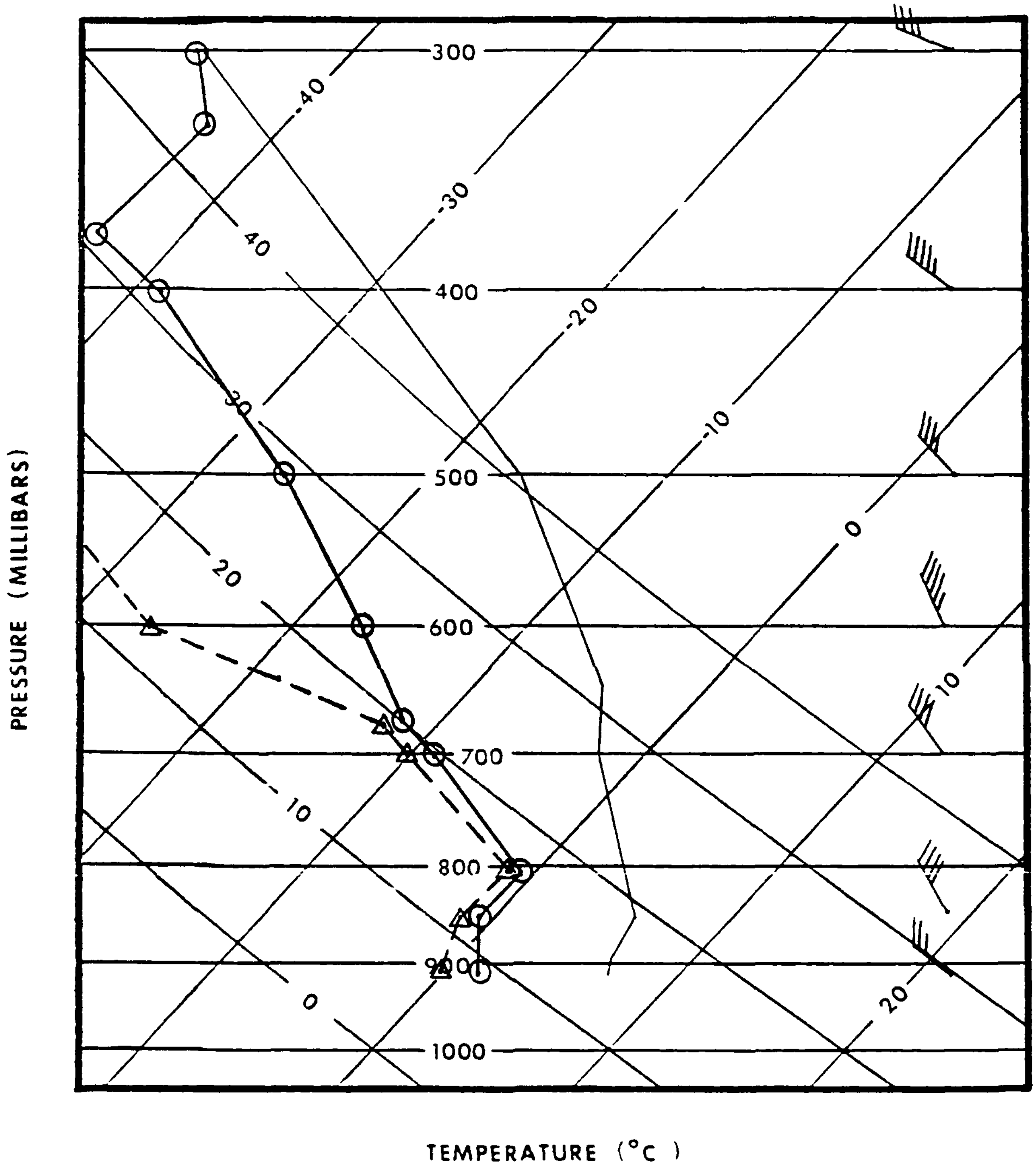


FIG. III-B-5 Dodge City, KS 72451 DDC for 1200 GMT 03 April 1984. The sounding shows cooling from the previous 24 h trace (the thin line to the right of the sounding). Note the cooling is deep, extending above the 400 mb level. The tropopause is at 376 mb. The lapse rate of the sounding is close to the wet adiabatic lapse rate. The mP air is above the 800 mb level, and the potential temperature at the bottom of the mP air is 16° C. Some colder air, probably cP is in the lower level. This is a typical mP front passage.

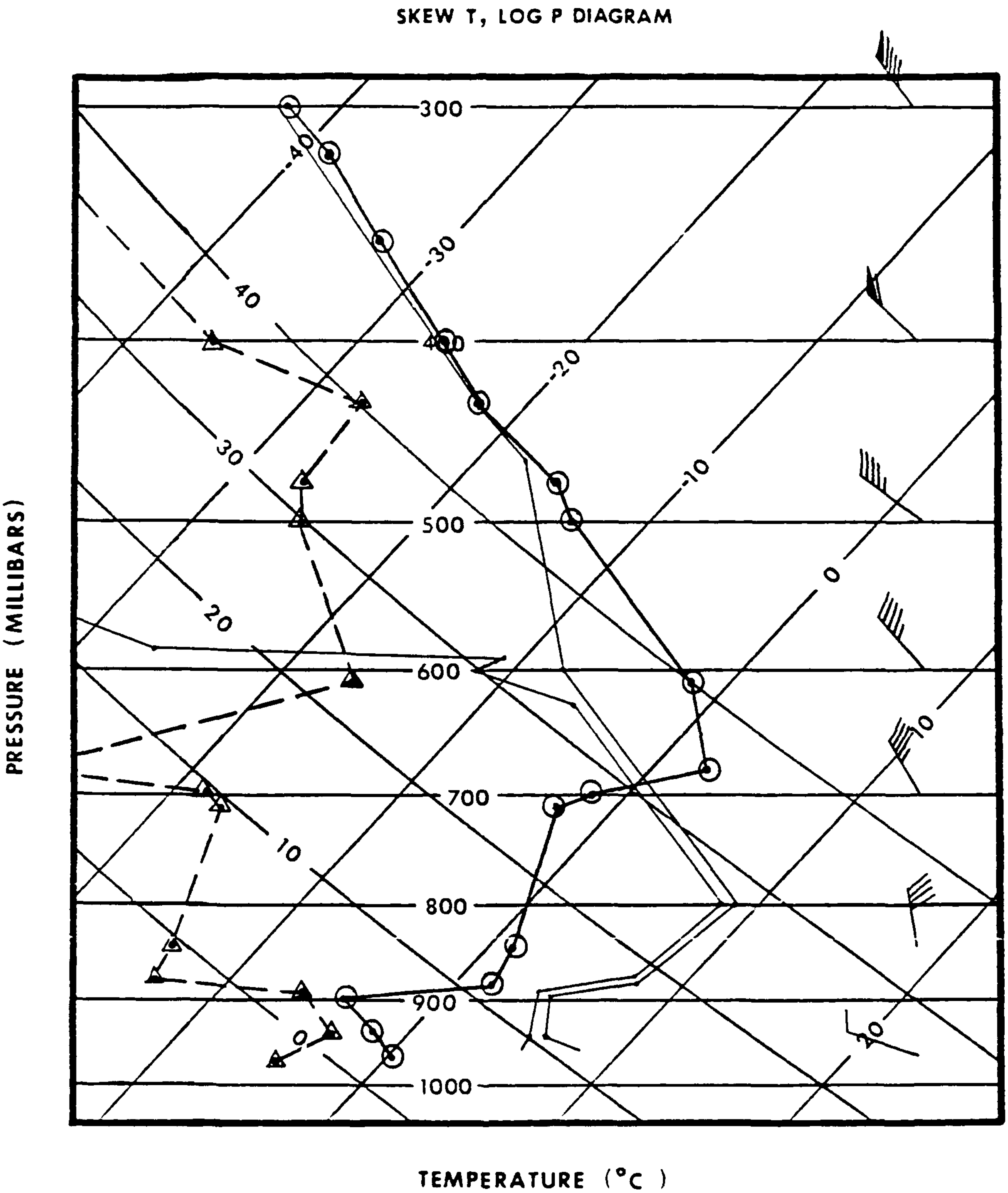


FIG. III-B-6 Oklahoma City, OK 72353 OKC for 1200 GMT 10 November 1983. This sounding represents a cP cold front passage. The top of the cP air is about 680 mb. Notice the cooling below the front and some warming above the front (24 h trace, the thin solid lines). The θ of the cP air is 11° C (at base). The air below 900 mb may be cA. The continuity of the cA front should be checked. Note the unstable air below 900 mb. This is caused by heating from the surface, which modifies the cA air rapidly.

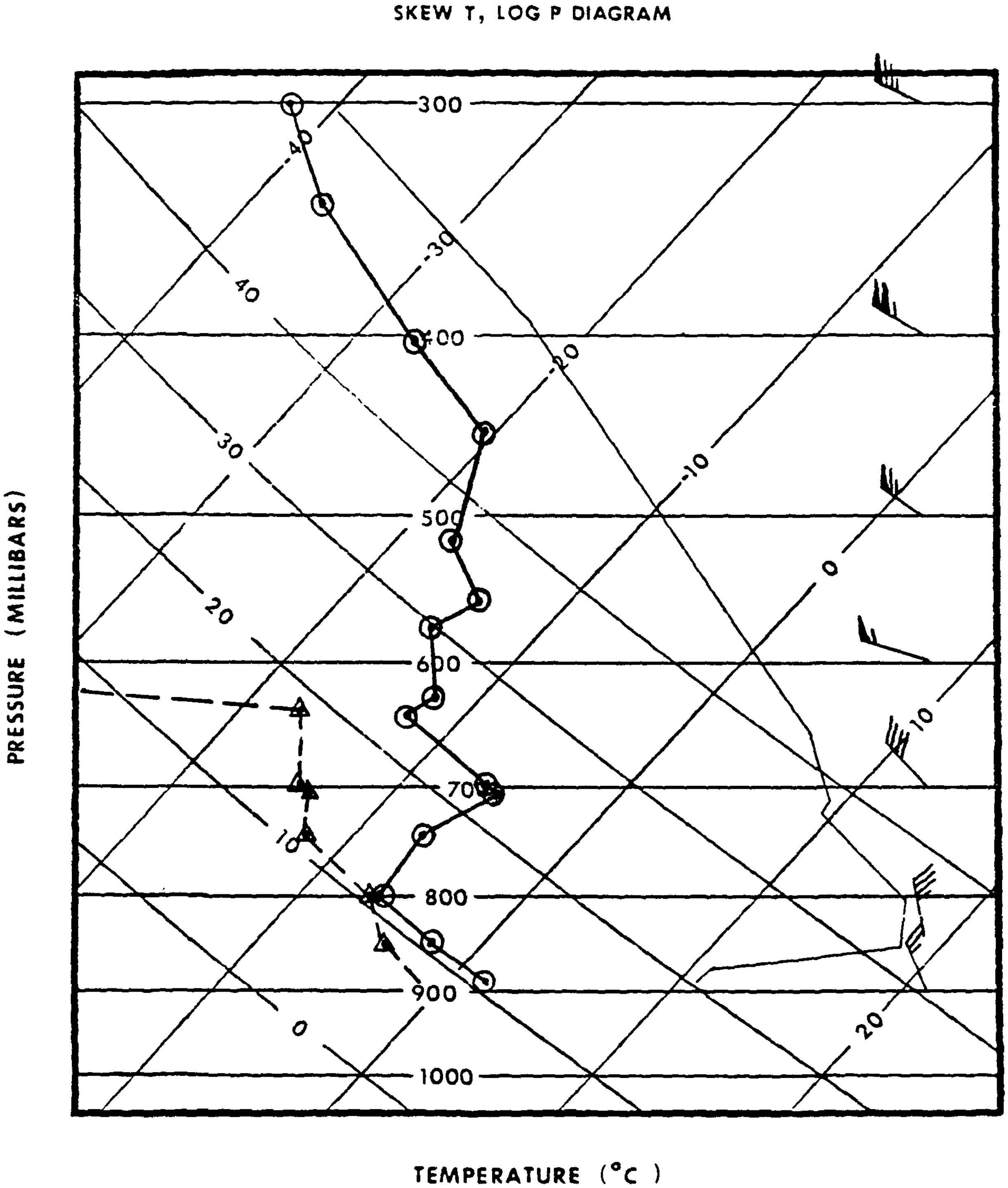


FIG. III-B-7 Amarillo TX 72363 AMA for 1200 GMT 09 November 1983. As so often happens the cP front catches the mP front and both pass the station, either at the same time or only a few hours apart. The sounding shows both fronts, the cP front from the surface to 700 mb and the mP aloft. The trace (thin line) shows cT air over AMA above 850 mb so the great cooling. Note the wind is backing. The unstable layer below 800 mb is caused by heat exchange with the surface.

The cP air may retreat as a warm front and be replaced by another air mass is shown in Fig. III-B-8.

In the United States, especially the midwest, the orientation of the front will be a guide to the type. The mP fronts tend to orientate NE-SW while the cP fronts are more E-W.

5. Frontal Analysis. Fronts are defined as the boundary between air masses. The surface and upper air masses must be determined. In synoptic analysis the front and air mass is checked on previous maps to determine the source region and the trajectory. Then it is declared to be mP, cP, cA, etc. Single station analysis does not permit the backtracking to the source region so the characteristics of the airmass need to be studied. The geographical location and season may limit the possibilities. The frontal analysis is easier and can be done with greater confidence when the station is in the cold air mass at the surface. The orientation of the front is determined by the orientation of the 1000-700 mb or 1000-500 mb thickness lines, which are parallel to the thermal wind of the layer. Since neither the thickness lines nor the thermal wind are available, the best approximation of the orientation and gradient of the thickness lines is the shear vector. Shear vectors should be computed for the layer from the GLW to 700 mb wind and from the GLW to 500 mb wind. When the fronts are shallow the GLW-700 mb wind shear should be more representative and for deep fronts the GLW-500 mb wind shear should be used. The orientation of the front will be about the same as the shear vectors. The magnitude of the shear vector will be proportional to the strength of the front.

The distance between the front and station can be determined when the station is in the cold air. The upper surface of the frontal zone should be located on a plotted Skew T-Log P Diagram. When attempting to locate the top of the front remember that the θ value will remain about the same on a time series of soundings and that the winds should show cyclonic shear in the frontal zone. The temperature advection as shown by the winds will indicate if the front is warm or cold. From the surface to the 850 mb level cP and cA warm fronts have a slope of about 1:250 and cold fronts about 1:50. Above the 850 mb level the fronts have a flatter slope. The slope of the front is determined by the temperature and wind shear across the front and some other factors (speed of move-

SKEW T, LOG P DIAGRAM

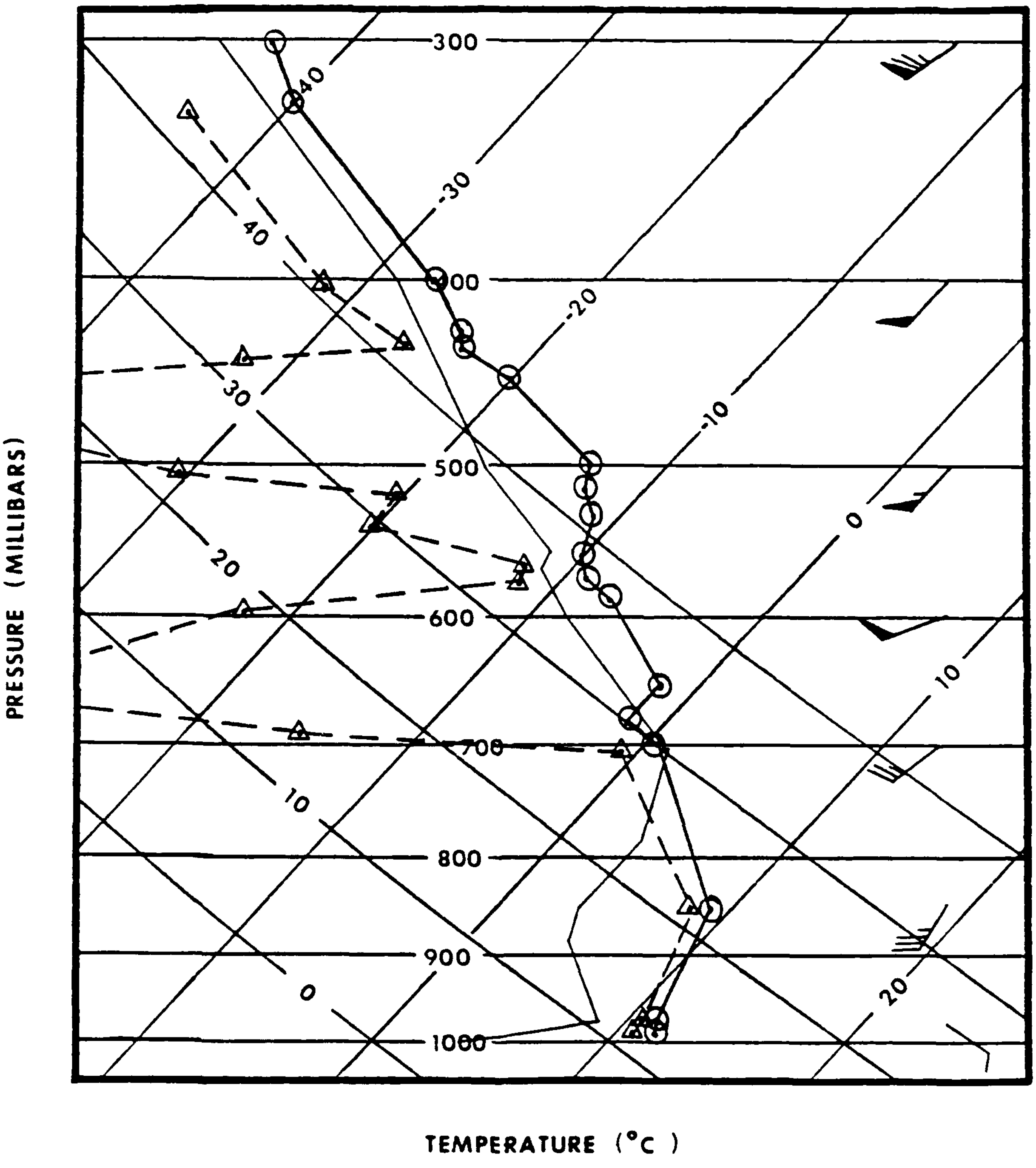


FIG. III-B-8 Nashville TN 72327 BNA for 1200 GMT 03 April 1984. This sounding represents a weak warm front passage. The trace (thin solid line) showed the front was at 700 mb but by 24 h the front had passed so that the cP air was replaced with mT. The surface Θ of 12° C will warm during the day to 23 or 24° C, thus better representing the mT air. The dry layer between 700 and 600 mb is caused by subsidence which often occurs when a warm front passes. Note the warm air advection indicated by the winds.

ment, latitude and type of air masses) so an adjustment to those slopes may be made when the front has added conditions. Measure the distance from the station and orientate the front parallel to the selected shear vector.

To compute the distance of a front use the following equation.

$$\text{Distance} = \frac{\text{Ht. Top of Front - Sta. Height}}{5280} \times \frac{(\text{Slope of Front})^{-1}}{69} \quad \text{(III-12)}$$

For most uses the heights may be read from the Pressure Altitude Scale. The 5280 converts the feet to miles and the 69 changes the distance to degrees latitude. Say the top of the front is 750 mb (8000 ft P.A.), station height 880 ft, and the slope is 1:50.

$$\text{Distance} = \frac{8000 - 880}{5280} \times \frac{(\frac{1}{50})^{-1}}{69} = 0.98° \text{ Lat.}$$

The same equation (III-12) can be used using the metric system. The equation then becomes

$$\text{Distance} = \frac{\text{Ht. Top of Front - Sta. Height}}{1000} \times \frac{(\text{Slope of Front})^{-1}}{111} \quad \text{(III-12)}$$

The 1000 converts the meters to km and the 111 changes the distance from km to degrees latitude.

Using the same data 8000 ft ≅ 2438 m and 880 ft ≅ 268 m.

$$\text{Distance} = \frac{2438 - 268}{1000} \times \frac{(\frac{1}{50})^{-1}}{111} = 0.97° \text{ Lat.}$$

Another method of estimating the distance to the front from the station is by timing the front. The cold front will move at a speed of about 80% of the component of the GLW normal to the front (the GLW must be in the cold air). A warm front moves at about 50% of the component of the GLW normal to the front (the GLW must be in the cold air). In a study of the wind and temperature structure of eleven cold fronts which passed an instrumented tower during the nighttime, Brundidge (1965) found that the speed of the front is well-represented by the component of the 600 ft wind normal to the front. Little error would result from using the 1000 ft wind. These results were based on measurements within the first several hours after passage. A warm or cold front can be extrapolated in time up to 12 h.

Use continuity, climatology and standard models to complete the map. Draw the frontal model and orientate it to fit the data. Continuity always helps, for example, if the winds were from the southwest before a cold front passed the station, the winds will still be from the southwest ahead of the cold front. If thunderstorms preceded the front they will continue to precede the front. Local climatology may dictate a diurnal change of intensity. Non-adiabatic conditions also change the characteristics of the front. Time of day, land forms, land-water, season, clear or cloudy sky all have an impact.

The maps should have vertical consistency or "stacking." If the maps do not "stack" hydrostatically then something is wrong with the basic analysis. On the upper air maps the isotherms and contours should show the warm and cold advection in the correct areas in relationship to fronts, troughs, ridges and other features. The isotherms should pack to the cold side of the front and the isotherm next to the front should be the temperature associated with the potential temperature of the front adjusted for the level.

6. The Stability Chart. With the radiosonde data several stability indices may be computed, see Grover (1979) or Miller (1972). In addition to the stability indices the location of more and less stable areas is important. The instability wind or shear of shears may be used for this purpose (see discussion III-A-4). The shear vector between the 850 and 700 mb winds is determined and called the first shear. Determine the shear vector between the 700 and 500 mb levels. This is the second shear. Now compute the shear between the first and second vector. This shear vector is called the shear of shears or the instability vector. The more stable area is to the right and the less stable area is to the left. An example of the calculation is shown in Fig. III-B-9.

In addition, the change of stability in the lower part of the atmosphere and the boundary layer can be computed as shown in III-A-4 (see Fig. III-A-3). Diurnal heating also becomes important so an estimate of the convective temperature T_c (see AWSTR 79/006) should be made.

PART C AN EXAMPLE

1. Data. The observations will continue from Chapter II. The surface observations for Station #3 are given in Table III-C-1.

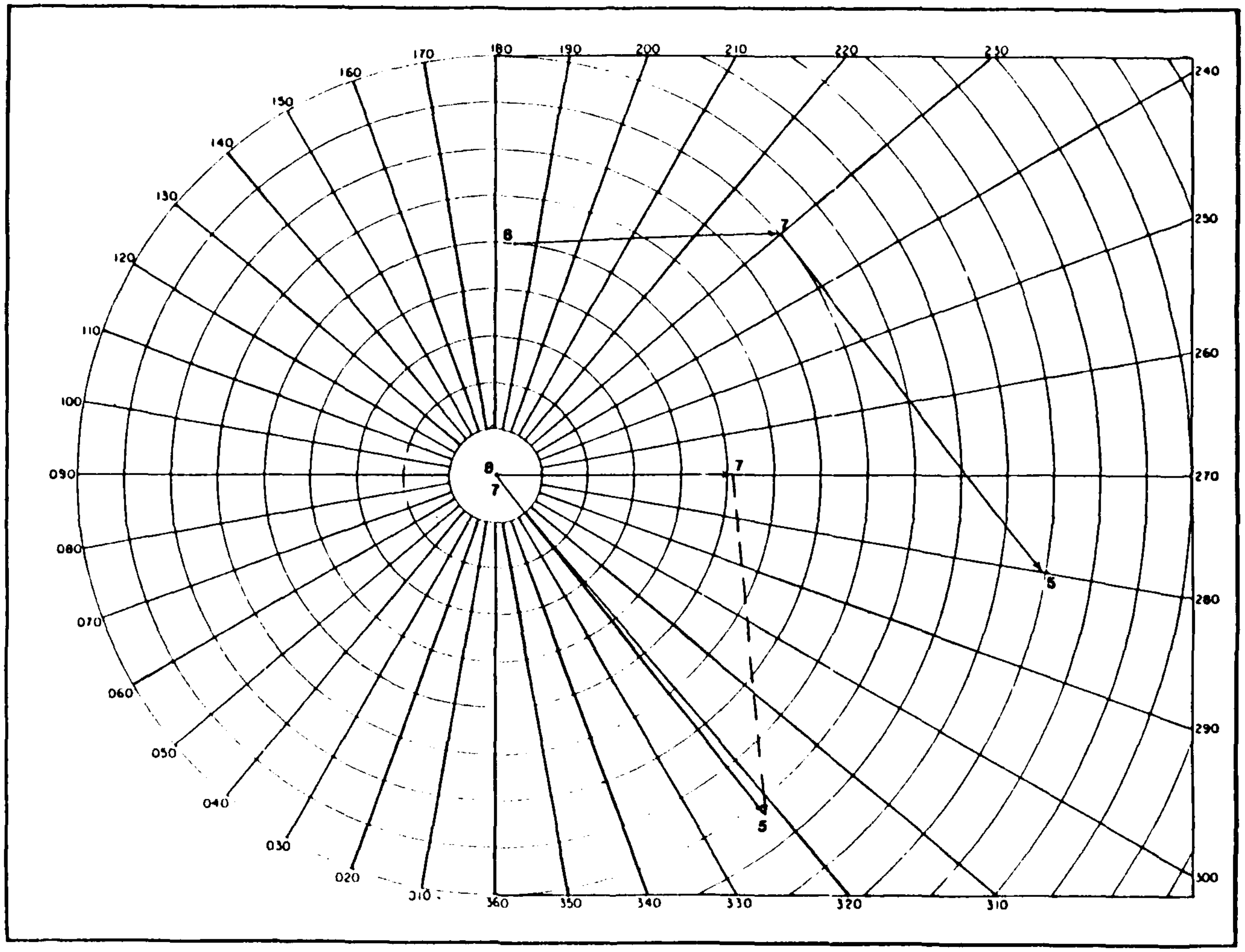


FIG. III-B-9 Computation of the shear of shears (instability vector). 850 mb wind 185°/25 kt, 700 mb wind 230°/40 kt, 500 mb wind 280°/60 kt. Resolve 850-700 mb shear to origin. Resolve 700-500 mb shear to origin. Dashed line is shear of shears, from 355°/38 kt. Unstable air is to the east.

TABLE III-C-1 SURFACE OBSERVATIONS FOR STATION #3 D3-0600 LST TO D3-1800 LST

DAY	TIME (LST)	SKY	VSBY (miles)	WW	SEA-LEVEL PRESSURE (mb)	TEMP (°F)	DEW POINT (°F)	RH	WIND DIR (°)	WIND SPEED (KT)	SKY DETAIL N_s	CC	$h_sh_sh_s$	N_s	CC	$h_sh_sh_s$	N_s	CC	$h_sh_sh_s$	REMARKS
D3	0554	E13 OVC	7+		123	32	29		315	20	10	SC	13							SC MVG SE 20KT
D3	0617	15 SCT							315	20	4	SC	13							
D3	0656	CLR	7+		145	28	19		315	18										
D3	0755	300 SCT	7+		156	24	15		340	14	4	CS	300							
D3	0854	140 SCT E300 OVC	7+		170	25	14		340	12	3	AS	140	10	CS	300				AS MVG E 40 KT
D3	0955	E150 BKN 300 OVC	7+		177	25	12		340	12	6	AS	150	10	CS	300				
D3	1055	E150 BKN 300 OVC	7+		184	25	10		360	9	7	AS	150	10	CS	300				
D3	1155	E120 BKN 300 OVC	7+		180	28	15		360	9	6	AS	120	10	CS	300				
D3	1255	120 SCT E300 OVC	7+		169	27	14		360	8	4	AS	120	10	CS	300				AS MVG E 40KT
D3	1357	E120 BKN 300 OVC	7+		159	28	15		360	9	6	AS	120	10	CS	300				
D3	1454	E120 BKN 300 OVC	7+		156	27	13		360	10	5	AS	120	10	CS	300				
D3	1556	120 SCT E300 OVC	7+		154	26	12		020	9	4	AS	120	10	CS	300				
D3	1657	120 SCT E300 OVC	7+		162	25	16		050	8	4	AS	120	10	CS	300				
D3	1755	120 SCT E300 OVC	7+		162	24	12		050	7	4	AS	120	10	CS	300				AS MVG E 40KT

2. Interpretation of the Data, D3-0600 to D3-1800 LST.

a. Low Clouds. The low clouds which were behind the front have moved on south and east so the sky has cleared.

b. Air Mass. The cold air mass (probably cP) is dry so no low clouds or fog will occur.

c. Advection. The continued cold air advection matches the solar heating and the temperature is not changing much. The drier air is being advected as indicated by the lowering dew point.

d. Wind Directions. The winds were from the northwest, shifted to northerly and shifted again to be from the northeast. This indicates that the high is moving eastward and passing north of the station. The easterly movement is consistent with the movement of the altostratus clouds, and with the recorded pressures.

e. High Clouds. Altostratus and cirrostratus clouds normally are associated with an approaching warm front. The extension of the cold front to the west may have become stationary or even be approaching as a warm front. The winds from the northeast are not favorable for continued southerly motion of the front because they are parallel to the front.

3. Winds Aloft. During D3 the upper air observing equipment arrived at Station #3 and was put into operation as soon as possible. A visual observation of the winds aloft was made using a theodolite and a balloon at D3-1235 LST. The observed winds are listed in Table III-C-2, and plotted on a hodograph in Fig. III-C-1.

TABLE III-C-2

Winds aloft for D3-1300 LST for Station #3.

Height ft MSL	Direction	Speed kt
sfc.	360°	08
1000	360°	08
2000	350°	06
3000	350°	07
4000	310°	07
5000	270°	12
6000	270°	15
7000	270°	18
8000	270°	25
9000	270°	31
10000	270°	33
12000	280°	39

Balloon lost in clouds.

a. Advection. The backing winds between the 3000 ft wind and 5000 ft wind indicates cold air advection.

b. Depth of cold air. The top of the cold air is at 5000 ft (top of cold air advection).

c. Stability. The very uniform winds from the surface to 3000 ft indicate mixing and a dry adiabatic lapse rate.

d. Frontal Orientation. The front would be orientated from 260° to 080° (shear vector computed using the 3000 ft wind to 10,000 ft wind).

4. D3-1800 LST Rawinsonde. The rawinsonde was launched to obtain the D3-1800 LST sounding. The winds and mandatory level data are given in Table III-C-3.

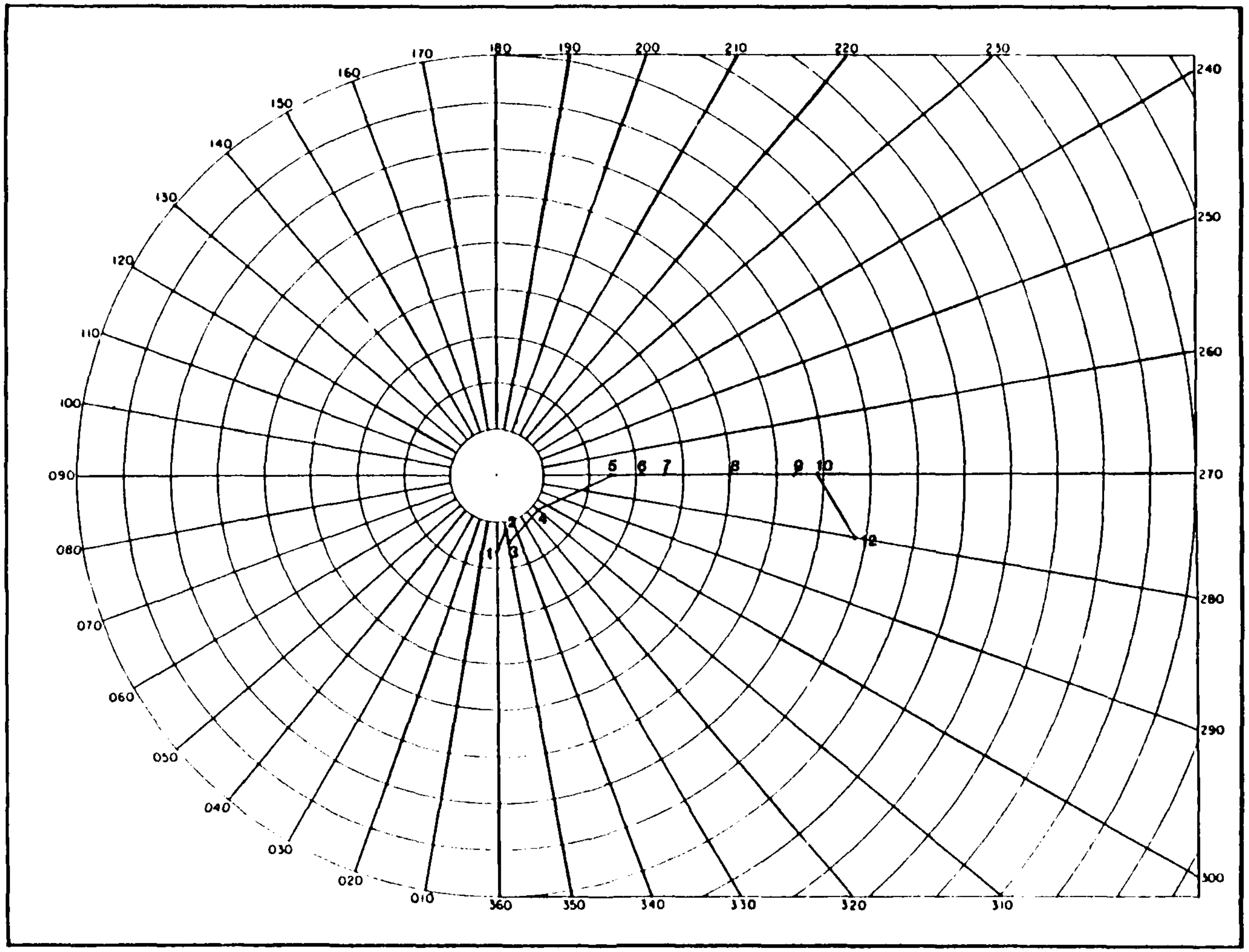


FIG. III-C-1 Station #3 PIBAL of D3-1300 LST.

TABLE III-C-3

Station #3 D3-1800 LST Rawinsonde Data

	Wind	TT	$T-T_d$	HHH
2000 ft	0710			
3000 ft	0511			
4000 ft	3512			
850 mb	3315	-1.2	16.0	414
6000 ft	2818			
8000 ft	2524			
700 mb	2525	-6.5	14.0	950
12000 ft	2635			
16000 ft	2548			
500 mb	2653	-22.2	4.0	550
20000 ft	2666			

5. <u>The Analysis of the D3-1800 LST Rawinsonde Data For Station #3</u>.

a. Winds. The first step is to plot the wind data on a hodograph, Fig. III-C-2a and 2b. Then plot the data at the station on the surface, 850, 700, and 500 mb maps. The data are shown in Fig. III-C-4, 5, 6, and 7. Add the GLW to the surface map by using another color (like purple). In Fig. III-C-4 it is shown with a longer shaft. Compute (on hodograph or calculator) the GLW-700 mb and the GLW-500 mb shear vectors and plot on the surface map (see Fig. III-C-2a and III-C-4). Usually they are almost the same direction and only one needs to be plotted on the map. If different directions occur then plot both for further consideration. In this case the difference in direction between the two vectors is 11°, which will not change the solution.

Review the example in Part II-C in Chapter II to establish the continuity. On D3-0300 LST a cold front passed and from the wind directions at surface (Table III-C-1) it should become stationary. The continuity is shown in Fig. III-B-1.

As can be seen on Fig. III-C-2a the wind component normal to the shear vectors (GLW-700 mb wind and GLW-500 mb wind) is small and indicates very weak cold air advection. At the standard levels the shear vector for the 850 mb chart indicates weak cold advection while the shears for both the 700 and 500 mb charts indicate weak warm air advection, see Fig. III-C-2b.

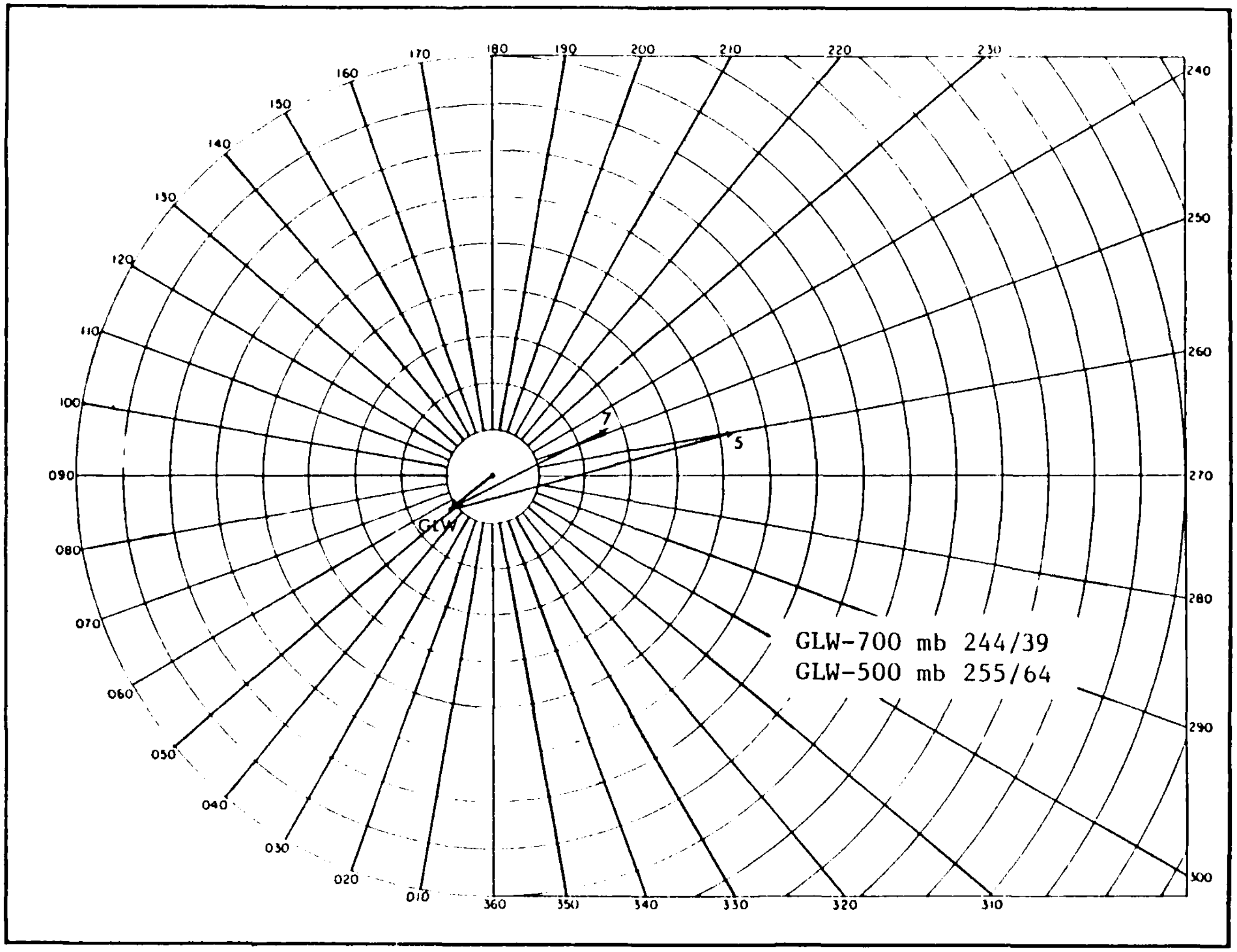


FIG. III-C-2a Shear vectors for the gradient wind (GLW) with the 700 mb wind and the 500 mb wind for D3-1800 LST for Station #3.

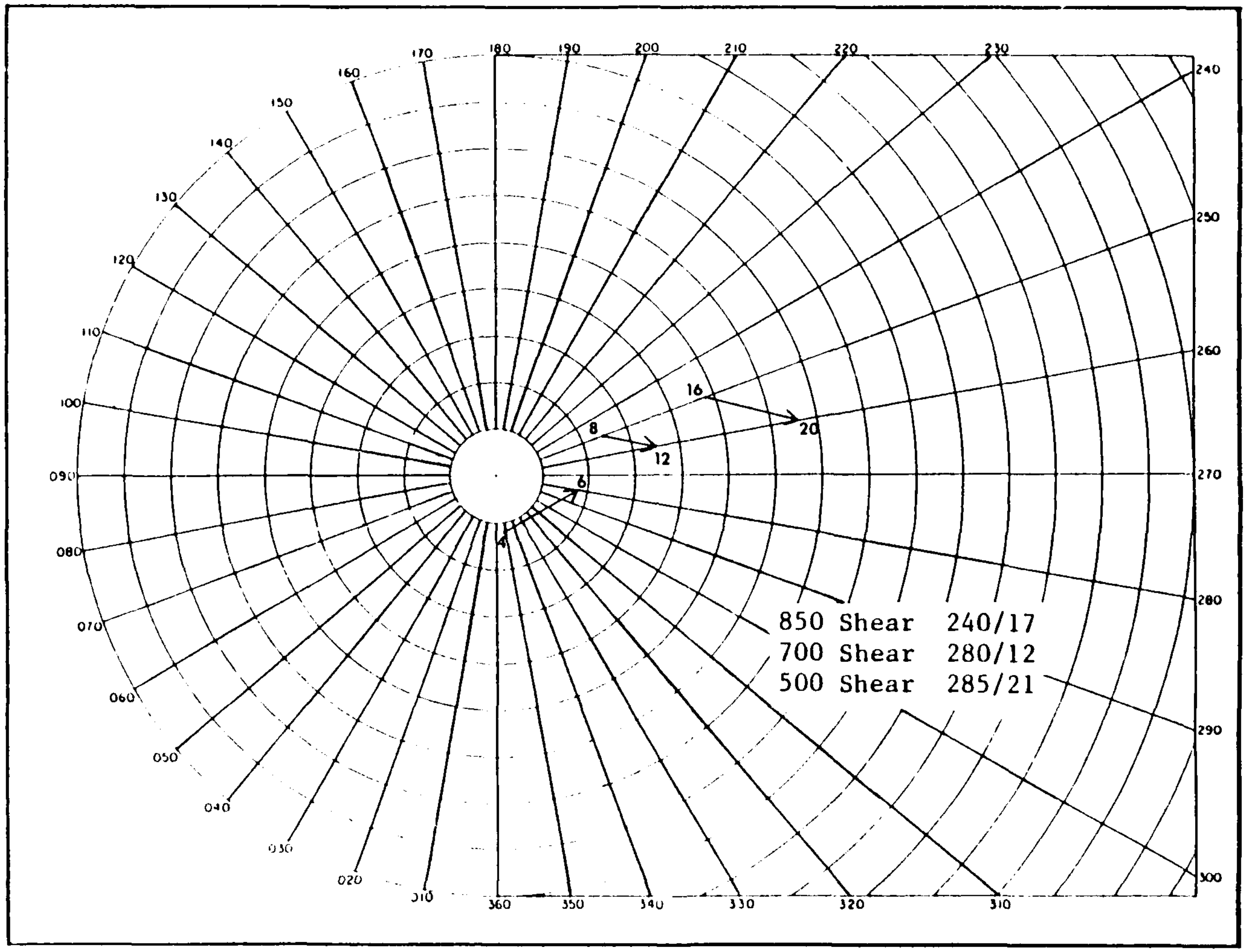


FIG. III-C-2b Shear vectors for the 850 mb, 700 mb, and 500 mb maps for D3-1800 LST for Station #3.

b. Front. The sounding (Fig. III-C-3) indicates that the top of the cold air is at 750 mb (8000 ft). The slope of the front is a function of the wind shear across the front (stronger--more vertical) and temperature difference across the front (greater ΔT--less slope). The horizontal wind shear across the front cannot be determined, but a strong wind shear in the vertical is an indication that the horizontal wind shear is strong. The shear of the GLW-700 mb wind is 39 kt which is sufficient thickness gradient for a moderate front. The temperature across the front is to be measured at the inversion, by using the difference of potential temperature across the front which is 24° C (22° C at top of inversion and -2° C at bottom). This ΔT while not the ΔT of Margules Equation is the best approximation of the ΔT across the front. Considering that the front is stationary with a good temperature gradient and a small wind shear, a slope of 1:200 would be about right since it is changing from the steeper slope of a cold front. In practice the heights may be read from the pressure altitude scale.

To compute the distance from the station see equation (III-12). In this case

$$\frac{8000 - 800}{5280} \times \frac{200}{69} = 3.9^\circ \text{ Lat.} \tag{III-12}$$

Since the GLW-700 mb wind vector is more representative of the cold air (the cold air is only 750 mb deep), the GLW-700 mb wind will be used as the guide to orientate the front. The front will be placed parallel to the GLW-700 mb shear and at a distance of 3.9° Lat with the station in the cold air.

c. Surface Map. The surface isobars may be drawn. Start at the station and place the 1014, 1016, and 1018 mb isobars using the GLW for orientation. From Appendix, Part I, the spacing of isobars at 2 mb interval for a wind speed of 11 kt at 40° N is 3.4° Lat. The distances are marked as heavy lines on the isobars on Fig. III-C-4. From the continuity the pressure at the front would be higher than it was at the time the front passed the station, because the low is moving northeastward so this section of the front is further from the low than the part of the front which passed Station #3. Therefore a pressure of below 1014 mb is a reasonable value. Also the 1014 mb

SKEW T, LOG P DIAGRAM

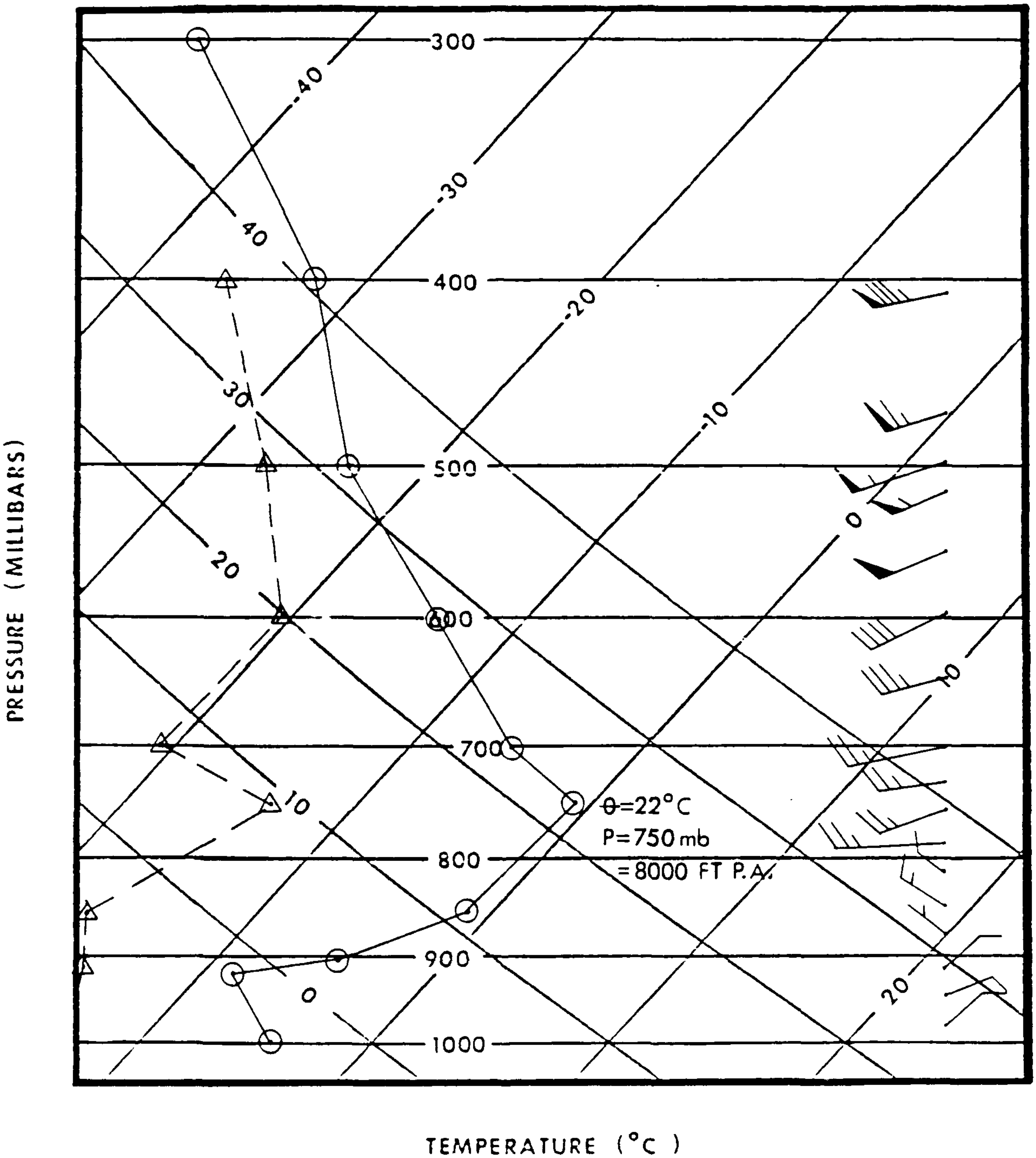


FIG. III-C-3 D3-1800 LST Station #3 plot of the lapse rate of T and T_d, and winds. The top of the front is at P=750 mb and θ is 22° C, and the height can be read from the pressure altitude scale. The base of the inversion is at 920 mb with a θ of -2° C.

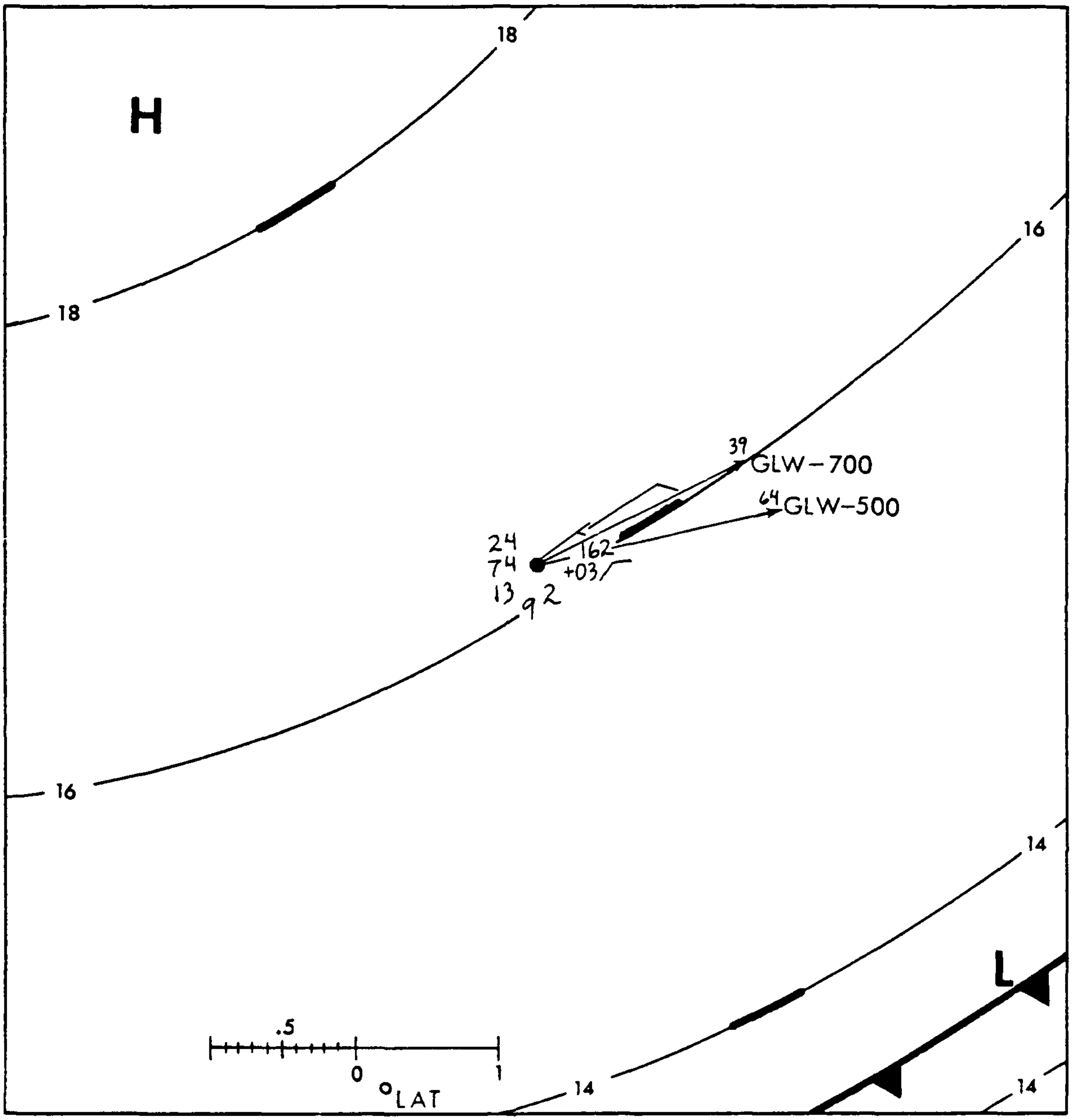


FIG. III-C-4 Surface map for Station #3 at D3-1800 LST. Thick sections of isobars indicate gradient spacing in relation to the gradient wind. Arrows are the shear vector winds and the number by the arrow head indicates the speed. Longer wind shaft indicates the gradient-layer wind.

isobar was turned away from the front toward the west indicating a stationary front. A low will probably form off of the southwest corner of the map.

d. 850 mb Map. Before the surface map was finalized the 850 mb map (Fig. III-C-5) should have been started. On the sounding (Fig. III-C-3) the 850 mb level is in the cold air, so the front is placed south of the station and parallel to the surface front. The contours are spaced using Appendix, Part J and the measurements are shown as heavy lines on the contours. Using the general pattern for fronts the low is to the northeast and also another low to the west or southwest so the 1440 contour is shown as turning away from the front to indicate the low. Before the front passed the winds at 850 mb were from the southwest and strong. The contours south of the front are drawn to maintain continuity. There is no reason to believe that the winds have not continued to blow from the southwest at about the same speed that they did prior to the front passing the station.

The shear vector for the 4000-6000 ft is shown on Fig. III-C-2b and is plotted on the 850 mb map, Fig. III-C-5. The isotherms are drawn parallel to the 4000-6000 ft wind shear and spaced by using the Appendix, Part L. The spacing of the isotherms is shown in heavy segments on the isotherms. The potential temperature of the top of the frontal inversion is 22° C and at 850 mb this is a temperature of 8° C. Thus a 8° C isotherm was drawn along the front in the cold air. The isotherms should pack on the cold side of the front. In Fig. III-C-5 the isotherms were spaced to fit the data at the station, then the temperature gradient tightened closer to the front and relaxed deeper into the air mass.

e. 700 mb Map. The top of the front is at 750 mb so at Station #3 the 700 mb level is in the warm air. The front is parallel to the surface and 850 mb front and is placed to show less slope (greater distance) at this level. (The distance from the 700 to 850 mb location of the front is greater than the distance from surface to 850 mb.) The contours are spaced by the wind and Appendix, Part J. At this level warm air advection is strong. To the north side of the front in the cold air the winds are still weak as compared with the winds in the warm air. The measured spacings of the contours are

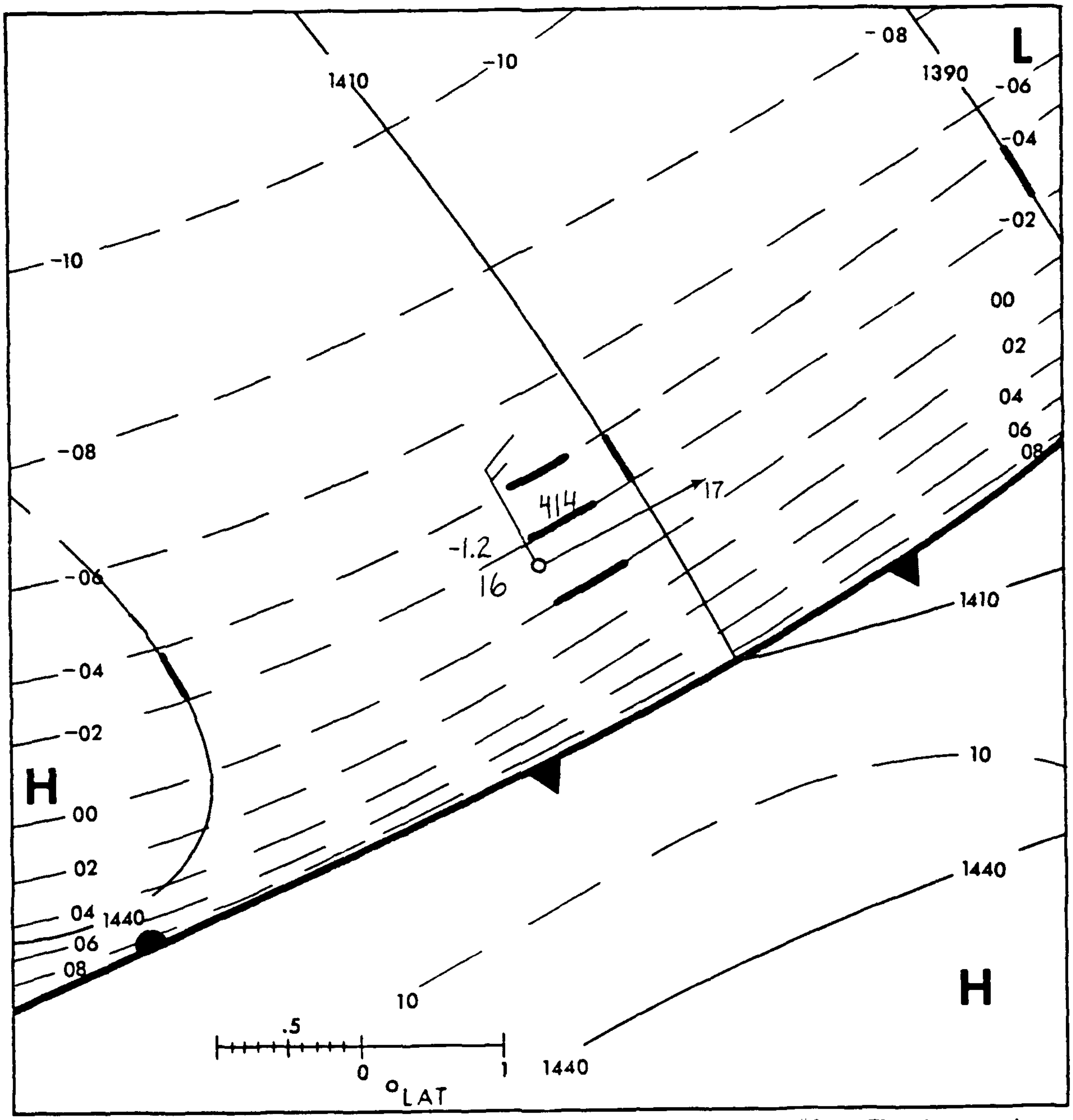


FIG. III-C-5 850 mb map for D3-1800 LST for Station #3. Thick sections on contours and isotherms indicate spacing. Isotherms are dashed and the arrow is the shear vector.

shown by the heavy marks on the contour lines. The isotherms are spaced using the 8000-12000 ft shear vector and the Appendix, Part L, and the spacing is indicated by the heavier section of the isotherm. The potential temperature of 22° C as used on the 850 mb map would be a temperature of -5° C at the 700 mb level. Thus the -5° C isotherm would be along the front at the 700 mb level. To the north side of the front the isotherms are drawn in a tight gradient, Fig. III-C-6.

f. 500 mb Map. The 500 mb map is shown as Fig. III-C-7. The spacing for the contours is read from the Appendix, Part J, and is shown by the heavy marks on the contours. The contours were given anticyclonic curvature. Refer to Fig. III-B-1c and notice that a weak ridge has passed the station 12 h previously. Therefore, the ridge should be drawn east of the station and the station is still in the ridge. This is an example of using the x-t chart for continuity. The isotherms are drawn parallel to the 16000 to 20000 ft shear vector (see Fig. III-C-2b) and their orientation relative to the height contours indicates warm air advection. The isotherms are spaced by using the Appendix, Part L. The map indicates a weak trough to the west, which matches with the low that was anticipated on the surface and the 850 mb maps. The warm air advection has caused a height rise of 60 m during the last 12 h.

g. Stability Map. The stability map, Fig. III-C-8, was constructed using the shear of shears and the graphical computation is shown on Fig. III-C-9. The Total-Totals (TTI), the Showalter (SI) and the "K Index" (KI) were computed and plotted at the station. These are all empirical indices and do not have any mathematical relationship to the shear of shears, thus, any spacing could be used. In this case all indices indicated stability. However, as an example, TTI isopleths were drawn to indicate the most stable air to the right of the vector, an area that is over the high on the 850 mb level which seems to be reasonable. However, note that the isopleths are curved to indicate more instability along the front. Maps at all levels and other supporting charts must indicate the same features if they are to be correct and useful.

h. The Surface Prognostic Chart (Fig. III-C-10). The front has become quasi-stationary to the south of the station. The GLW is parallel to

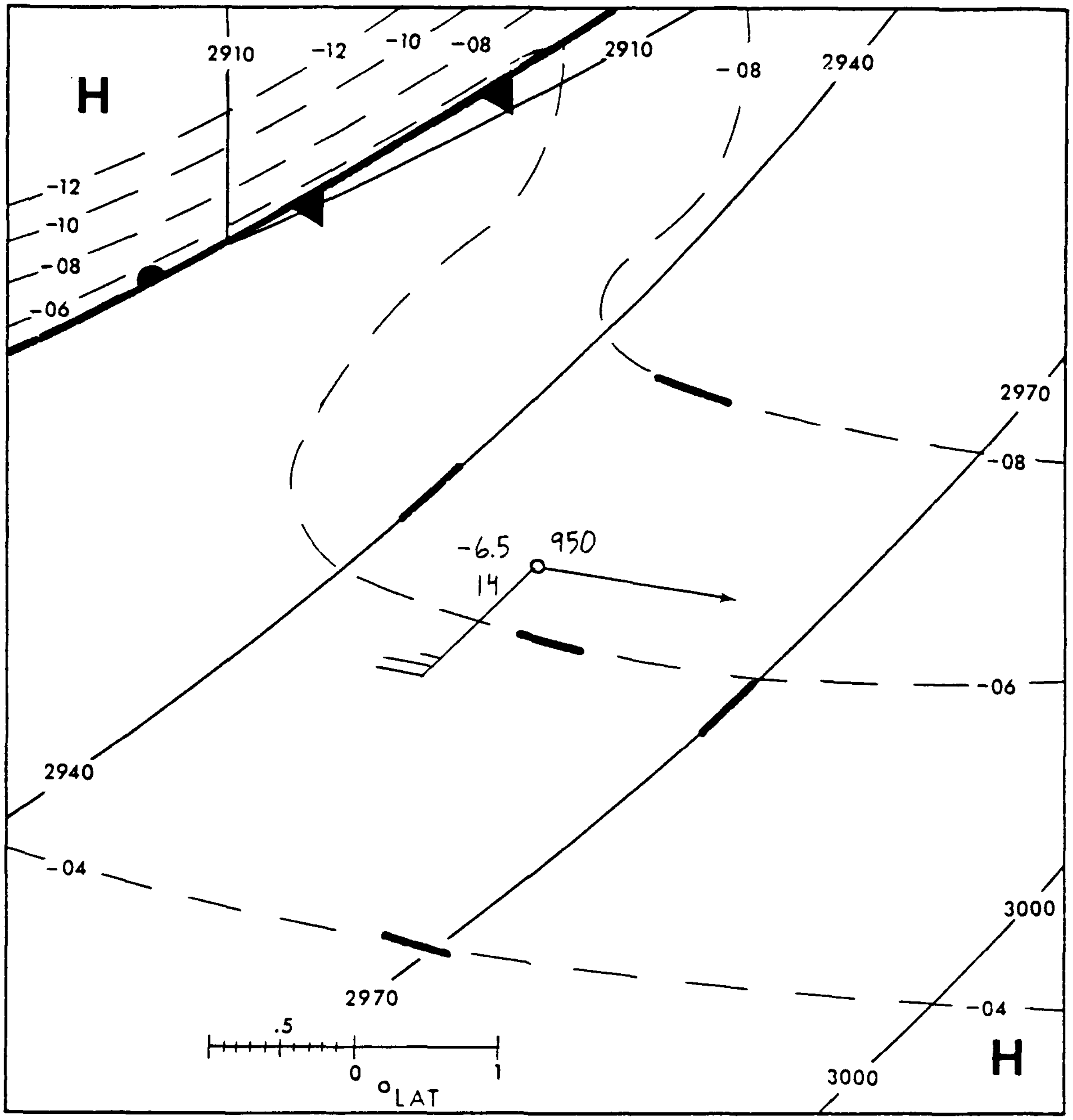


FIG. III-C-6 700 mb map for D3-1800 LST for Station #3. Thick sections on contours and isotherms indicate spacing. Isotherms are dashed and the arrow is the shear vector.

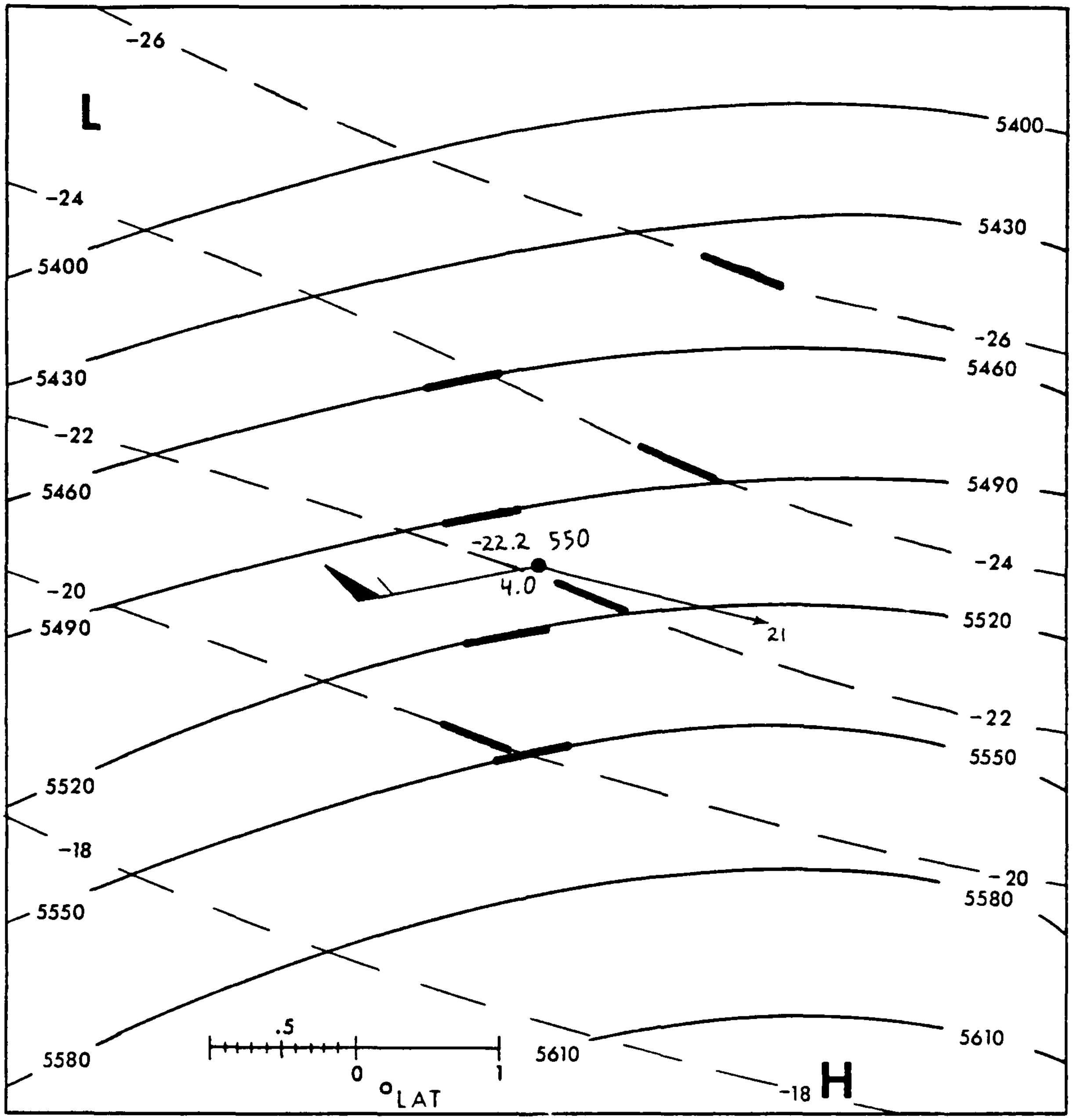


FIG. III-C-7 500 mb map for D3-1800 LST for Station #3. Thick sections on contours and isotherms indicate spacing. Isotherms are dashed and the arrow is the shear vector.

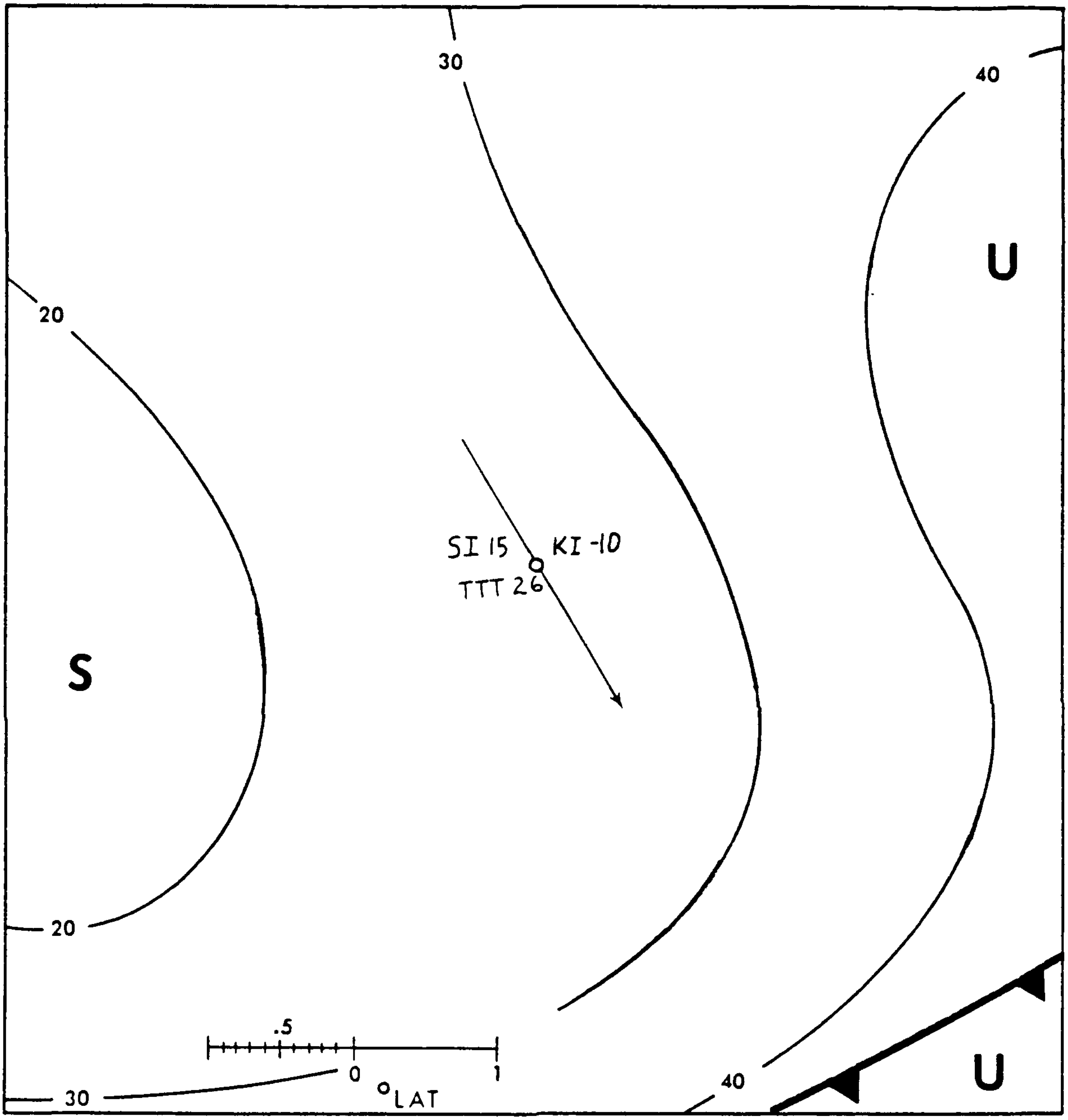


FIG. III-C-8 Stability map for D3-1800 LST for Station #3. The arrow is the Shear of Shears and the S indicates more stable and the U more unstable.

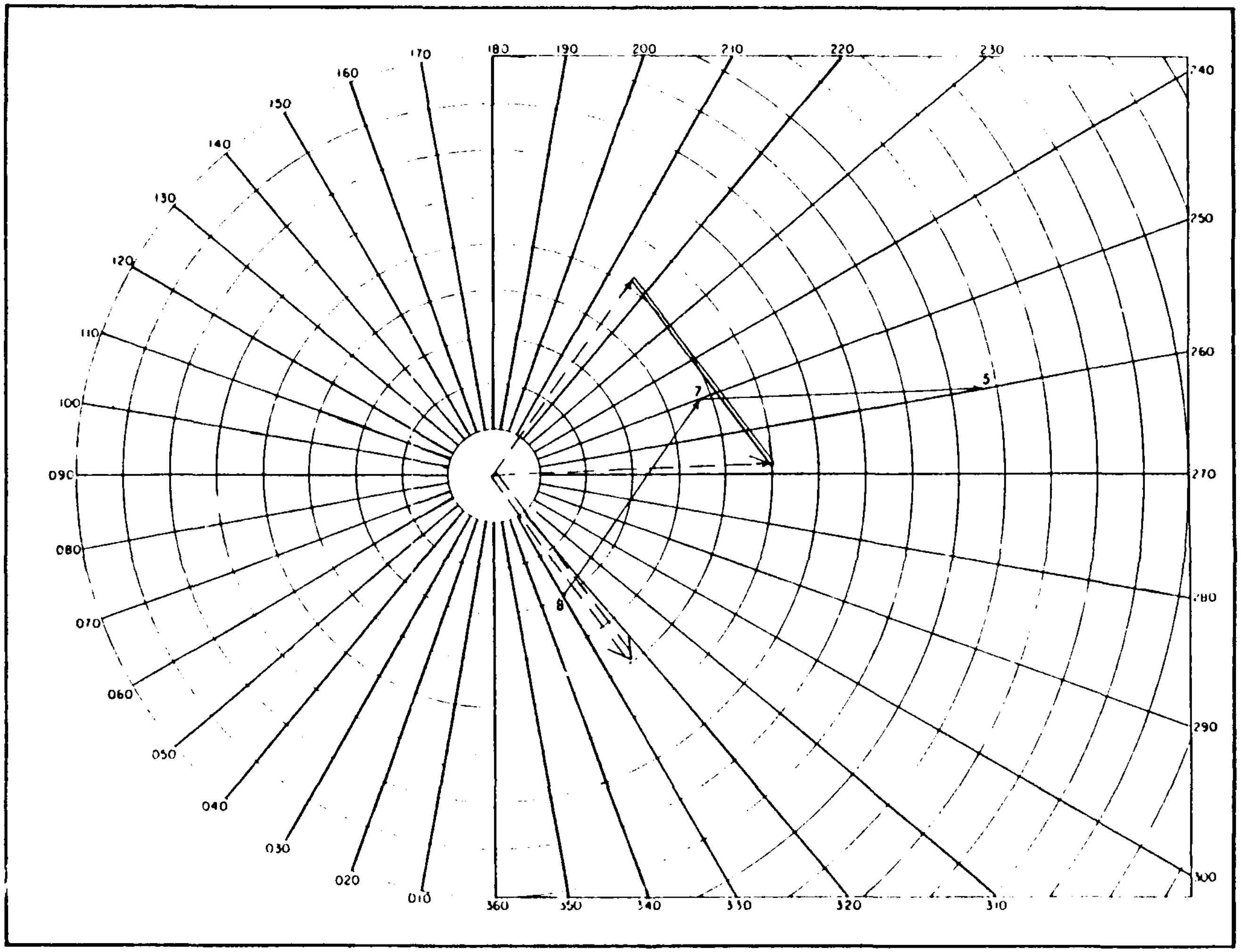


FIG. III-C-9 Computation of shear of shears for Station #3 for D3-1800 LST. Shear of Shears is computed using the 850, 700, and 500 mb winds. The shear of the 850 and 700 mb is determined and resolved to the origin (the dashed line close to 220) and the same is done with the 700 to 500 mb winds (dashed line along 270). The Shear of Shears then is resolved to the origin and is 326° / 25 kt.

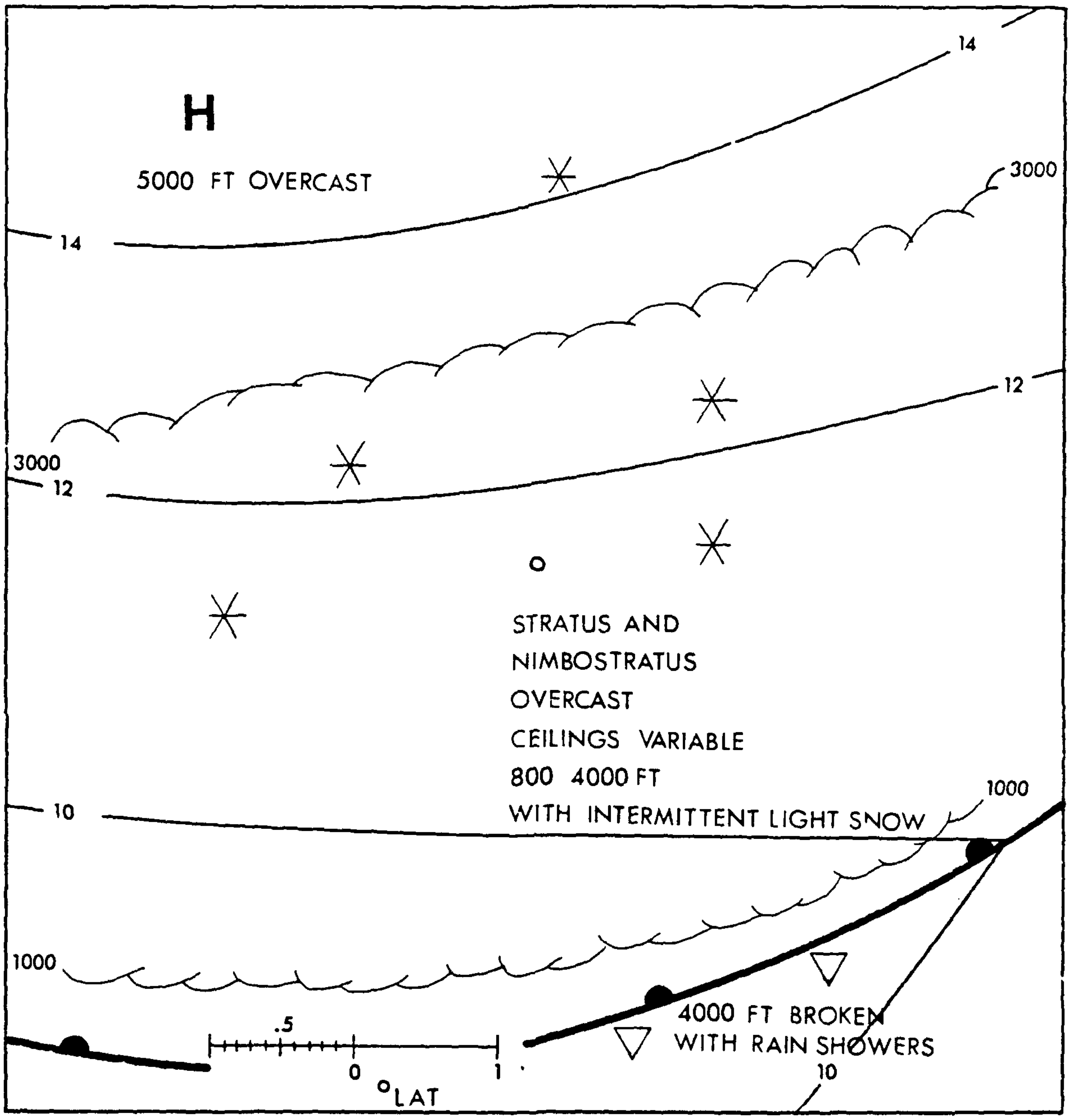


FIG. III-C-10 Surface Prognostic Map. V.T. D4-0600 LST for Station #3.

the front so little movement of the front would be expected during the first 6 h. However, with the low off the west side of the map the front should become a warm front by D4-0000 LST, and start a slow movement northward; the pressure should fall as the low moves eastward. The major trough should remain to the west off of this map.

i. Weather. Winds above the front are from the southwest which are over-running the warm front so clouds will lower with precipitation. Consider the warm front in the standard wave and place the weather and clouds. (Refer to Fig. II-B-1a, II-B-1b, and II-C-1.)

1.) the present condition. 120 SCT E300 OVC 7+ 24/12/0507.

2.) sky condition. Ceiling will lower to 1000 ft overcast with temporary ceilings to 800 ft. Higher ceilings of 5000 ft north of the station (this is based on the over-running warm front condition).

3.) visibility. 7+ with occasional reduction in snow.

4.) weather. Light snow starting after midnight. No significant accumulation.

5.) temperature. No change; will remain in the mid 20's° F.

6.) winds. The wind will be from the northeast 10-15 knots all period. Equivalent chill temperture 0° F.

7.) south section (area south of the front). Ceilings 4000 ft with rain showers, especially along front (showers when front passed Station #3); winds from the southwest at 15 knots and temperature in low 60's° F (temperature at #3 prior to cold front with some more warm advection).

8.) outlook. Conditions will remain about the same for the next period 0600 LST to 1800 LST, turning colder with increasing snow tomorrow night as the low to the west moves eastward.

CHAPTER IV

THE SPARCE DATA PROBLEM

PART A THEORY OF UTILIZING SPARCE DATA

1. Introduction. The techniques for dealing with the single-station forecast problem have been discussed in the foregoing chapters. However, there may be situations in which a region involving several weather stations may be isolated, or otherwise may not have access to the observations from outside the region. The recommended procedure in this case is to first apply SSA techniques to each station and then combine the information to complete the analysis. An example will be given in this chapter by the addition of observations from two more stations to those of Station #3.

With observations from as many as three stations, it is possible to make computations of quantities that cannot be determined with observations from a single station. The following will give the theory for and examples of the computations of horizontal divergence, vertical motion, vorticity, and computed change of the Lifted Index. The latter includes methods for making short-term forecasts of temperature and moisture change.

2. Vertical Motion. With observations from three stations it is possible to calculate the vertical motion in the region enclosed by a triangle whose vertices are the given stations. Theoretically, upward motion should occur in the region between the trough and the downstream ridge in the 500 mb height field and cloud layers and precipitation may result. Conversely, subsidence should occur between the trough and the upstream ridge.

While the omega equation (Holton, 1979) would be the most desirable relationship for this purpose it cannot be used because the gradient of vorticity cannot be determined with data from only three stations. Another possible approach is to use the adiabatic method; however, this requires knowing the local change of potential temperature at the time of the observations. An estimate of this quantity based on the change during the past 12 h is the only method available, which makes its validity questionable. A third method, which is still subject to criticism but recommends itself on the basis of simplicity, is the assumption of incompressibility. One relationship for this computation may be written as

$$\int \nabla \vec{C} \cdot dV = 0 \ , \qquad \text{(IV-1)}$$

where $\vec{C}$ is the three dimensional velocity vector, $dV = dxdydz$, and the integration is over the volume bounded on the sides by the legs of the triangle and at the bottom and top by surfaces of one's choosing, e.g., the local terrain and 5000 ft, respectively. Applying the divergence theorem to (IV-1) gives

$$\int \hat{n} \cdot \vec{C} \, d\sigma = 0 \ ,$$

from which

$$(\overline{w}_T - \overline{w}_B)A' \cong - \int\int V_n dsdz \quad .$$

Here, the bars indicate an average over the area of the triangle, A'. The subscripts, T and B, designate the vertical motion at the top and bottom surfaces of the volume, V_n is the component of the wind normal to a lateral boundary of the volume and dsdz represents an increment of area on a lateral boundary. The integral on the right hand side can be approximated by a summation to give

$$\overline{w}_T = \overline{w}_B - \frac{1}{A'} \sum_{i=1}^{3} (\overline{V}_n)_i A_i \qquad \text{(IV-2)}$$

where $(\overline{V}_n)_i$ is the average normal component of the wind over the i^{th} lateral face of the volume and A_i is the area of the i^{th} face.

If α is the direction of the wind as reported in meteorology, i.e., the direction from which the wind is blowing, then the velocity components are given by

$$u = - V \sin \alpha \qquad \text{(IV-3)}$$
$$v = - V \cos \alpha$$

where V is the reported speed at a point. The computation of $\hat{n} \cdot \vec{V}$ is made by treating the sides of the triangle as if they were vectors which, like the wind, are from a direction given by the angle β as measured in meteorological convention (see Fig. IV-A-1). The unit vector $\hat{n}$ must be a vector rotated 90° clockwise from the direction of the side and such that $\hat{n}$ is directed outward from the triangle, as shown in Fig. IV-A-1. Then

$$\hat{n} = \hat{i}(-\cos\beta) + \hat{j}(\sin\beta).$$

Therefore, $$\overline{\hat{n} \cdot \vec{V}} = \overline{v} \sin\beta - \overline{u} \cos\beta = \overline{V}_n \ .$$

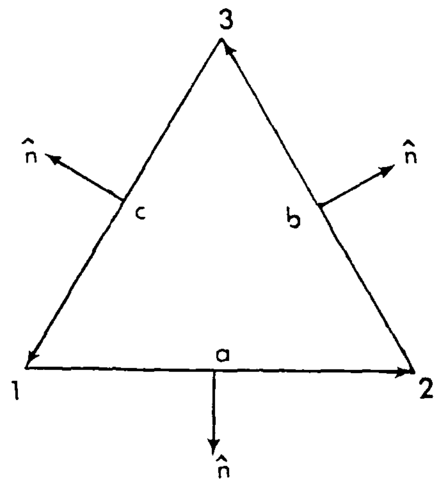


FIG. IV-A-1 Three-Station Triangle. Numbers indicate locations of the three stations. Letters identify sides of the triangle which, in turn, are represented by arrows to show their orientation in the meteorological system.

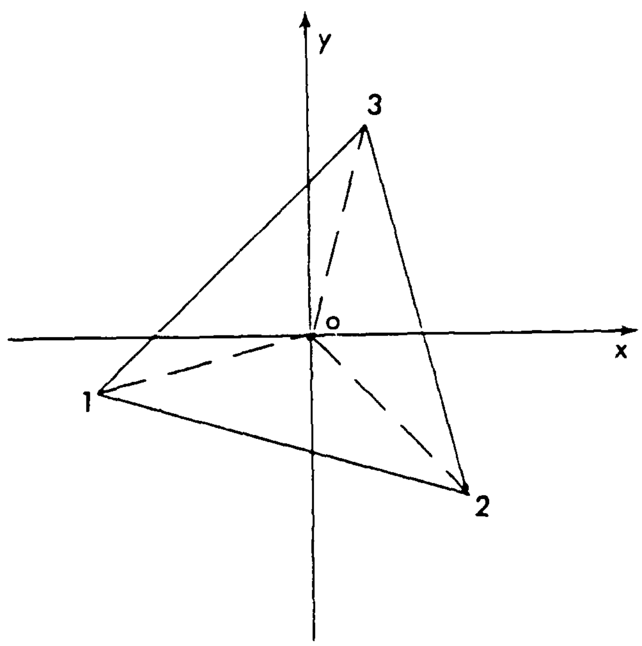


FIG. IV-A-2 Station Triangle in x, y Coordinate System. Numbers show locations of stations. The origin of the x, y coordinate is placed at the centroid of the triangle.

By this convention, outflow from the triangle is +, signifying divergence while inflow, or convergence is −. Note that $\bar{u}$ and $\bar{v}$ are simple arithmetic averages over a given face. For example, for face a in Fig. IV-A-1 $\bar{u}_a$ is given by

$$\bar{u}_a = \frac{(u_1)_B + (u_1)_T + (u_2)_B + (u_2)_T}{4}$$

Therefore, using the notation in Fig. IV-A-1, (IV-2) can be written:

$$\bar{w}_T = \bar{w}_B - \frac{1}{A'}[(\bar{v}_a \sin\beta_a - \bar{u}_a \cos\beta_a)A_a + (\bar{v}_b \sin\beta_b - \bar{u}_b \cos\beta_b)A_b + (\bar{v}_c \sin\beta_c - \bar{u}_c \cos\beta_c)A_c] \quad . \qquad \text{(IV-4)}$$

Subscript a refers to side a, and subscript b and c refer to sides b and c, respectively. A_a is the area of the vertical wall on side a, for example. By this method, the computation requires $\bar{w}_B = 0$ at the ground, in accordance with the kinematic boundary condition. Also, it requires fairly accurate knowledge of the terrain heights along the legs of the triangle in order to get accurate values ^f the A_i. An example is shown in Chapter IV-C-8.

An alternative relationship to (IV-2) comes from vertical integration of

$$\nabla \cdot \vec{V} + \frac{\partial w}{\partial z} = 0 \quad .$$

Thus,

$$w_T = w_B - \left(\frac{\nabla \cdot \vec{V}_T + \nabla \cdot \vec{V}_B}{2}\right)\Delta z \quad . \qquad \text{(IV-5)}$$

The subscripts in (IV-5) refer to two levels (T, top and B, bottom) separated in the vertical by the geopotential height increment, Δz. Values for $\nabla \cdot \vec{V}$ can be obtained from wind values at the three stations, as explained below.

If Q is some quantity whose values are known at the vertices of a triangle, then these values can be expressed by a set of linear Taylor Taylor Series expansions. Thus,

$$Q_1 = Q_o + (\frac{\partial Q}{\partial x})_o x_1 + (\frac{\partial Q}{\partial y})_o y_1, \qquad \text{(a)}$$

$$Q_2 = Q_o + (\frac{\partial Q}{\partial x})_o x_2 + (\frac{\partial Q}{\partial y})_o y_2, \qquad \text{(IV-6) (b)}$$

$$Q_3 = Q_o + (\frac{\partial Q}{\partial x})_o x_3 + (\frac{\partial Q}{\partial y})_o y_3. \qquad \text{(c)}$$

In these equations the subscripts 1, 2, 3 refer to the values at the vertices of the triangle, as shown in Fig. IV-A-2. The centroid of the triangle is taken as the location of the origin of the x, y coordinate system and the locations of the stations at the vertices must be assigned x, y coordinate values relative to the centroid. Q_o, $(\partial Q/\partial x)o$, and $(\partial Q/\partial y)o$ are values at the centroid; they can be obtained by simultaneous solution of the set (IV-6) to give

$$Q_o = \frac{Q_1(x_2y_3-x_3y_2) + Q_2(x_3y_1-x_1y_3) + Q_3(x_1y_2-x_2y_1)}{\Delta}, \qquad \text{(a)}$$

$$(\frac{\partial Q}{\partial x})_o = \frac{y_1(Q_3-Q_2) + y_2(Q_1-Q_3) + y_3(Q_2-Q_1)}{\Delta}, \qquad \text{(b)}$$

(IV-7)

$$(\frac{\partial Q}{\partial y})_o = \frac{Q_1(x_3-x_2) + Q_2(x_1-x_3) + Q_3(x_2-x_1)}{\Delta}, \qquad \text{(c)}$$

$$\Delta = y_1(x_3-x_2) + y_2(x_1-x_3) + y_3(x_2-x_1) . \qquad \text{(d)}$$

When (IV-7) is applied to the velocity components at the three stations, then

$$u_o = \frac{u_1(x_2y_3-x_3y_2) + u_2(x_3y_1-x_1y_3) + u_3(x_1y_2-x_2y_1)}{\Delta}, \qquad \text{(a)}$$

$$v_o = \frac{v_1(x_2y_3-x_3y_2) + v_2(x_3y_1-x_1y_3) + v_3(x_1y_2-x_2y_1)}{\Delta}, \qquad \text{(b)}$$

$$(\frac{\partial u}{\partial x})_o = \frac{y_1(u_3-u_2) + y_2(u_1-u_3) + y_3(u_2-u_1)}{\Delta}, \qquad \text{(c)}$$

(IV-8)

$$(\frac{\partial v}{\partial y})_o = \frac{v_1(x_3-x_2) + v_2(x_1-x_3) + v_3(x_2-x_1)}{\Delta}, \qquad \text{(d)}$$

$$\left(\frac{\partial v}{\partial x}\right)_o = \frac{y_1(v_3-v_2) + y_2(v_1-v_3) + y_3(v_2-v_1)}{\Delta}\ , \tag{e}$$

$$\left(\frac{\partial u}{\partial y}\right)_o = \frac{u_1(x_3-x_2) + u_2(x_1-x_3) + u_3(x_2-x_1)}{\Delta}\ , \tag{f}$$

and $\nabla\cdot\vec{V} = \left(\frac{\partial u}{\partial x}\right)_o + \left(\frac{\partial v}{\partial y}\right)_o$.

If level B in (IV-5) represents the ground, then it is necessary in this procedure to obtain w_B, the vertical motion forced by the motion of the air over the terrain. This can be computed from

$$\begin{aligned} w_B &= \vec{V}_o\cdot\nabla z \\ &= u_o\left(\frac{\partial z}{\partial x}\right)_o + v_o\left(\frac{\partial z}{\partial y}\right)_o\ , \end{aligned} \tag{IV-9}$$

where the subscript refers to values at the centroid of the triangle. If Q in (IV-7) now refers to elevation, z, above sea level, the terms in (IV-9) can be evaluated using

$$\left(\frac{\partial z}{\partial x}\right)_o = \frac{y_1(z_3-z_2) + y_2(z_1-z_3) + y_3(z_2-z_1)}{\Delta}\ , \tag{a}$$

(IV-10)

$$\left(\frac{\partial z}{\partial y}\right)_o = \frac{z_1(x_3-x_2) + z_2(x_1-x_3) + z_3(x_2-x_1)}{\Delta}\ , \tag{b}$$

Then (IV-5) becomes

$$w_T = w_B - \frac{\Delta z}{2}\left[\left(\left(\frac{\partial u}{\partial x}\right)_o + \left(\frac{\partial v}{\partial y}\right)_o\right)_T + \left(\left(\frac{\partial u}{\partial x}\right)_o + \left(\frac{\partial v}{\partial y}\right)_o\right)_B\right] \tag{IV-11}$$

Using either method described here, the w calculated at the top of a given layer becomes the w_B for the base of the next layer. An example of the second method is shown in Chapter IV-C-8.

3. Vorticity. The relative vorticity at the centroid of the triangle can be calculated for any level from the relationships (IV-8 e and f) by substitution into the statement

$$\zeta_o = \left(\frac{\partial v}{\partial x}\right)_o - \left(\frac{\partial u}{\partial y}\right)_o\ . \tag{IV-12}$$

The absolute vorticity is given by

$$\zeta_A = f_o + \zeta_o\ ,$$

where f_o is the coriolis parameter at the latitude of the centroid. ζ_A generally is + but ζ_o may be + or −. If $\zeta_A > f_o$, then the cen-

troid may be located in a trough but also could lie in relatively straight flow with cyclonic shear, e.g., on the north side of a jet-stream maximum. If $\zeta_A < f_o$, the centroid may be in a ridge or in relatively straight flow and on the south side of a jet maximum. A sample computation is shown in Chapter IV-C-8.

4. Stability Change. For a single station the calculation of the local temperature change, of necessity, must neglect the individual changes of temperature experienced by parcels undergoing the adiabatic process during vertical displacement. However, as previously shown, with three stations a calculation of vertical motion can be made as an average over the area of the triangle formed by the stations. Furthermore, with three stations it is not necessary to rely upon the geostrophic assumption to obtain horizontal temperature advection. Finally, it also is possible to determine stability change while accounting for advection and radiation effects during the daylight hours.

It is convenient to predict the stability change in terms of a stability index, such as the Lifted Index. This index is determined (see Duffield and Nastrom, 1984) by first determining the pressure-averaged mixing ratio and temperature for the lowest 100 mb of a sounding. Secondly, a parcel having these average values is lifted while following an adiabatic process from a point 50 mb above the surface to its LCL, and thence, moist-adiabatically to 500 mb. If T' is the lifted parcel's temperature at 500 mb, then LI is defined as

$$LI = T_{500} - T' \quad .$$

The approach used here is to predict new values of T_{500} and T'. The predictions are based on the rawinsonde information at the three stations assuming there are no severe topographic features between the stations, such as a mountain or a deep valley. This restriction makes it possible to place greater reliance upon the techniques described earlier for estimating vertical motion. Also, it is assumed for computations of a forecast LI that no front exists between the stations or is likely to move through during the forecast period. These assumptions restrict the use of this technique to situations in which the three stations remain in the same air mass during the forecast period.

For the prediction of T_{500}, consider the relationship for the individual derivative and assume adiabatic changes of state to obtain

$$\frac{\partial T}{\partial t} = -\overline{V}\cdot\nabla T - w(\Gamma_d - \Gamma) \quad . \qquad \text{(IV-13)}$$

Γ_d and Γ are the dry adiabatic and observed lapse rate, respectively. The local derivative in (IV-13) can be approximated by a finite difference. Therefore, the forecast value T_t, of T_{500} is given by

$$T_t = T_o - [u_o(\frac{\partial T}{\partial x})_o + v_o(\frac{\partial T}{\partial y})_o + w_{500}(\Gamma_d - \Gamma)]\Delta t \quad . \qquad \text{(IV-14)}$$

In (IV-14), w_{500} can be calculated from either (IV-4) or (IV-11), u_o and v_o and the derivatives of T are obtained from (IV-8 a and b) and (IV-7 b and c).

In the lowest 100 mb, the change of the mean temperature is a consequence primarily of advection and the modification by radiation. Thus,

$$\frac{\partial \overline{T}}{\partial t} = (\frac{\partial \overline{T}}{\partial t})_{rad} - \overline{V}\cdot\nabla\overline{T} \quad .$$

The forecast value of $\overline{T}$ is then given by

$$\overline{T}_t = [\overline{T}_o + (\frac{\partial \overline{T}}{\partial t})_{rad}]\Delta t - [u_o(\frac{\partial \overline{T}}{\partial x})_o + v_o(\frac{\partial \overline{T}}{\partial y})_o]\Delta t \quad . \qquad \text{(IV-15)}$$

Once again, (IV-7) and (IV-8) should be used to evaluate the derivatives and the wind components. For this purpose, the centroid GLW can be used.

Unless precipitation occurs it is not unreasonable to assume that the air parcels will essentially conserve their mixing ratio values over the forecast period in the lowest 100 mb. Also, vertical advection is negligible in this layer and turbulent transport is ignored. Therefore, the forecast equation for the mean mixing ratio, $\overline{r}$, is

$$\overline{r}_t = \overline{r}_o - [u_o(\frac{\partial \overline{r}}{\partial x})_o + v_o(\frac{\partial \overline{r}}{\partial y})_o]\Delta t \quad . \qquad \text{(IV-16)}$$

When T_t , $\overline{T}_t$, and $\overline{r}_t$ are plotted on a Skew T-Log P Diagram, a predicted value of LI can be obtained in accordance with its definition.

The procedure just outlined cannot be demonstrated using data from the three stations considered in this chapter because of the presence of a front in the region. The use of the linearized Taylor-series expansions for temperature and mixing ratio requires that smooth gradients of these variables exist in the region; they actually have first-order discontinuities when a front is present.

PART B PRACTICAL PROCEDURES AND METHODS

1. Sparce Data. After working with only one station as described in Chapter III, any scrap of additional data becomes very valuable. The analysis can

be extended in the direction of the one report with confidence. Also, only one of the stations should be surprised by a weather event because the others would be forewarned.

If a network of three stations can be established the forecaster may have three times the amount of data but the advantages are cubed. The geographical relationship of the stations does make a difference. An equilateral triangle is the best but a line may be very useful. Any arrangement is better than not having the data. A fourth or fifth station continues to improve the situation. In this study three stations will be used. The addition of more stations increases the number of possible triangles or extends the line, but the methodology does not change.

2. Adding Data. With sparce data the single station procedures of Chapters II and III should be continued and applied to all stations available. Plot data at all stations for all levels.

   a. Frontal Analysis. Continuity of fronts and the passage of the fronts as indicated by the hourly data should be identified and rate-time-distance relationships maintained. The depth of the cold air on each radiosonde should be noted. The potential temperature should be close to the same value (within 2 or 3° C) to identify the front. Do not mix fronts or inversions of different potential temperature values. The distance from the station to the surface frontal positions can be determined for each sounding as in Chapter III-B-5. These locations of the front from all three stations should be mutually supportive or the forecaster needs to rethink the work. A tentative position of the fronts can be placed at the various pressure levels, especially the surface and 850 mb charts.

   b. Contour Analysis. In Chapter III-C the contours were spaced in relation to the calculated distance for the given GLW speed and maybe some correction for curvature added after a study of the z-t chart and the determination that a ridge or trough was diagnosed to be in the area. With two or three stations the forecaster will notice that when the real winds are treated as if they were geostrophic in order to obtain contour spacing there may be difficulty in drawing a consistent pattern. Now is the time to consider curvature. The gradient wind equation (III-3) is composed of the geostrophic wind and a curvature factor. If there is cyclonic curvature the wind will be weaker than the geostrophic wind and, if anticyclonic curvature, the wind will be faster than the geostrophic wind. Study Fig. IV-B-1a, 1b, and 1c, which represent these conditions on a 500 mb chart. When only three heights are available, one is not eliminated by calling it incorrect. It must be rechecked and reconsidered before deciding it is wrong. The same concept works at all levels.

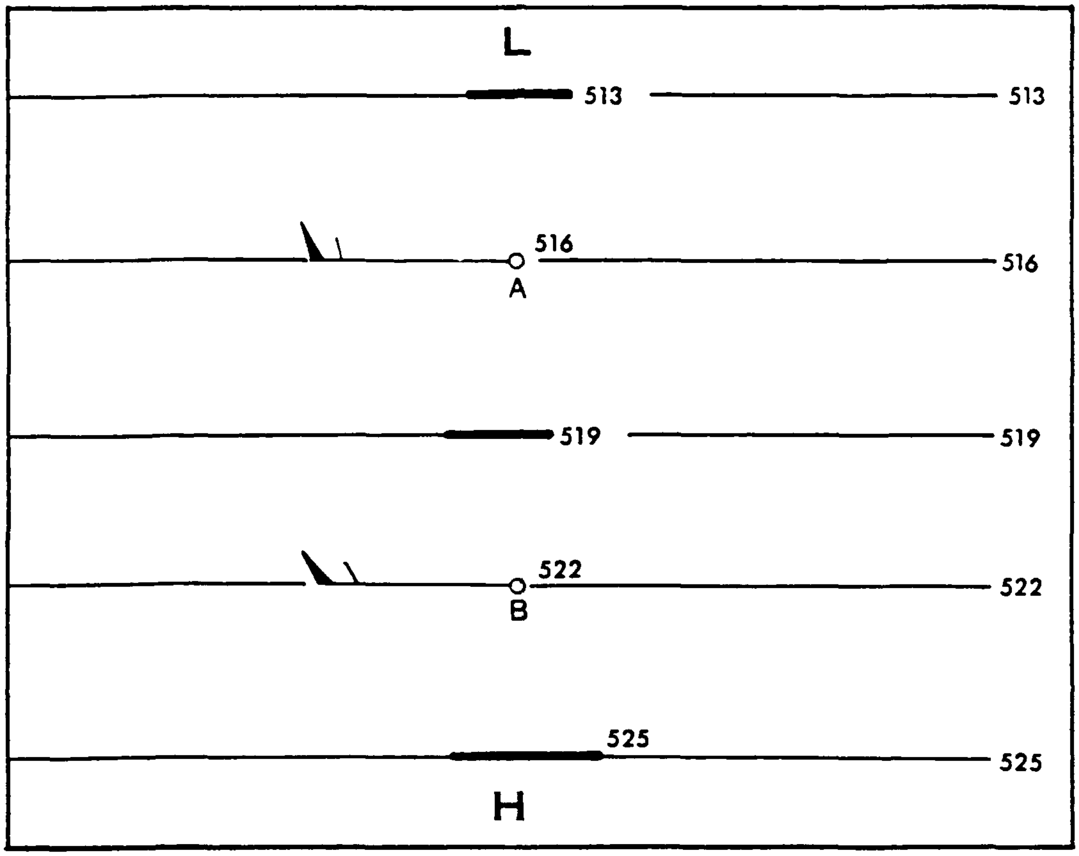


FIG. IV-B-1a 500 mb chart with straight contours. The spacing for the wind speed and the heights fit. The heavy marks on the contours indicate the measured spacing based on assuming the observed wind is in geostrophic balance.

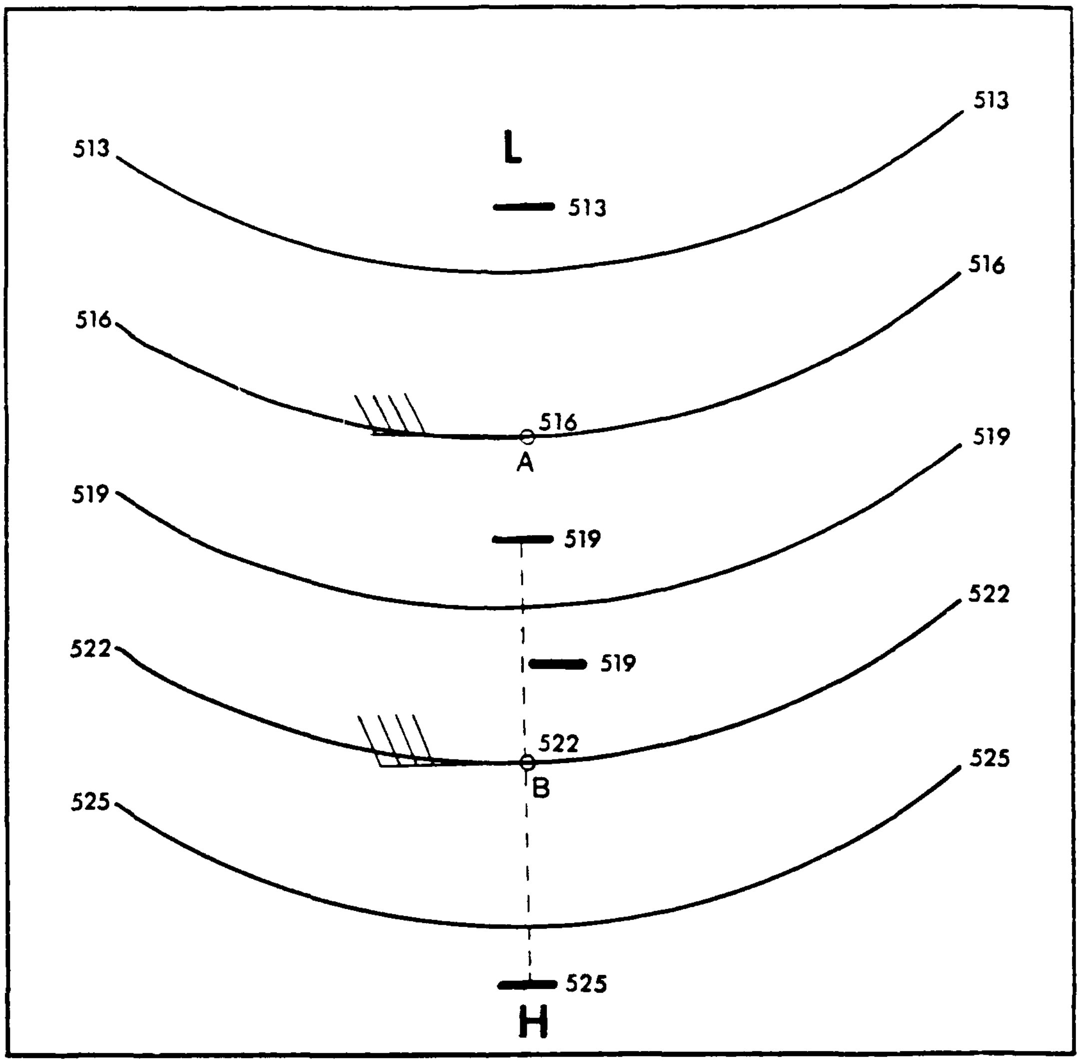


FIG. IV-B-1b 500 mb chart with cyclonic contours. The winds are less than geostrophic in speed, so the geostrophic spacing is too wide to fit the height data. The two heavy lines on either side of Station A and marked 513 and 519 represent the spacing required for a geostrophic wind. The same distance is marked from Station B, labeled 519 and 525 and connected with a dashed line through Station B. The 519 contour cannot go through both locations marked 519. However if the contours have cyclonic curvature the gradient may be tightened, and the pattern should be drawn as shown, which fits the heights, windspeed and curvature.

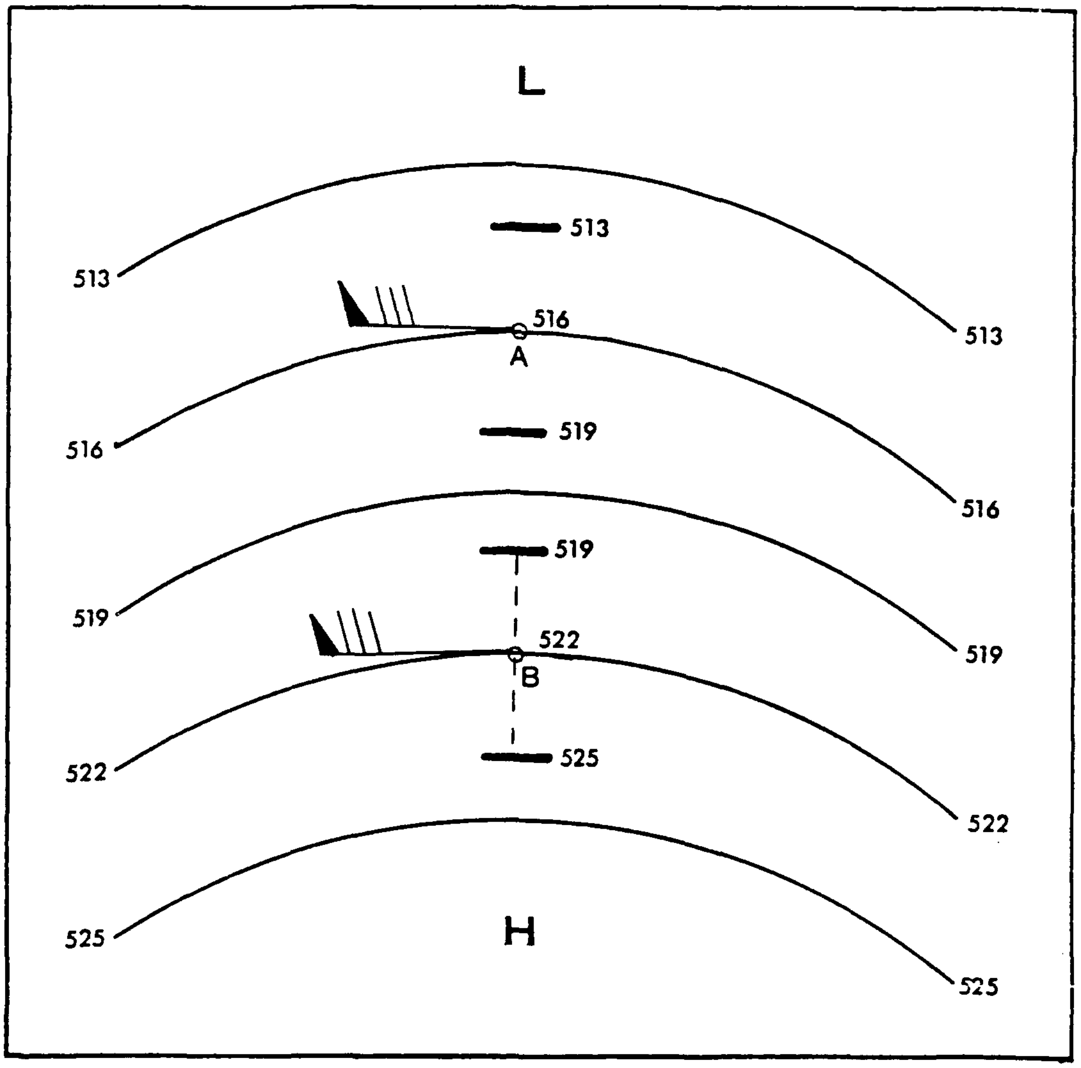


FIG. IV-B-1c 500 mb chart with anticyclonic contours. The winds are larger than geostrophic in speed so do not agree with the heights at the stations. The geostrophic spacing is indicated for Station A by the heavy lines at 513 and 519. The spacing for Station B is shown by the heavy lines at 519 and 515 and connected with the dashed line through Station B. The 519 contour cannot pass through both the locations for the 519. However if anticyclonic curvature is occurring the spacing would be greater and the contours could be drawn as shown.

The stations do not need to have the same wind speed or the same wind direction and do not have to be on a line perpendicular to the winds as shown in Fig. IV-B-1. Figure IV-B-2 shows a more realistic situation. The heavy line segments indicate the contour spacing using the observed wind as geostrophic. The necessary adjustments are indicated in Fig. IV-B-2.

c. Stability Analysis. The use of three stability vectors (shear of shears) will define the unstable areas better than one vector will. In the event the wind is not reported at some of the levels it may be estimated from the analyzed upper air chart, then the stability vector may be computed. As soon as any map is analyzed winds at all points can be estimated, and additional stability vectors may be computed for the rest of the map if desired.

d. Moisture Analysis. The forecaster can make a moisture analysis on the upper air maps. The analysis may be of the $T-T_d$, r, T_d, or any other moisture variable desired.

1.) 850 mb moisture. On the 850 mb chart, and especially if showers or thunderstorms are expected, a T_d or r analysis would be most useful. Remember that at any pressure the T_d represents a r value so the isolines may be labeled as T_d at one end and r at the other. Continuity must be maintained because the water vapor (mixing ratio) will move with the wind and remain constant unless the water content is changed by evaporation, mixing with another layer or by condensation. Thus, a reasonable mixing ratio (dew-point) analysis may be made at 850 mb. Moist and dry intrusions may be identified. Do not analyze the isodrosotherm to be warmer than the isotherm.

2.) cloud analysis. On the 700 mb and 500 mb charts the $T-T_d$ values often are analyzed because then the chart can be used as a cloud chart. Extrapolation of relative humidity is not easy because of the vertical motion. However, when attempting to make such a chart refer to the cloud observations at the stations, the vertical motion as computed in Chapter IV-A-2 and the frontal patterns, and good estimates may be made.

e. Weather. With three stations reporting the weather and by moving the weather events with the steering winds aloft, a good idea of the weather in and about the triangle or line of stations can be maintained.

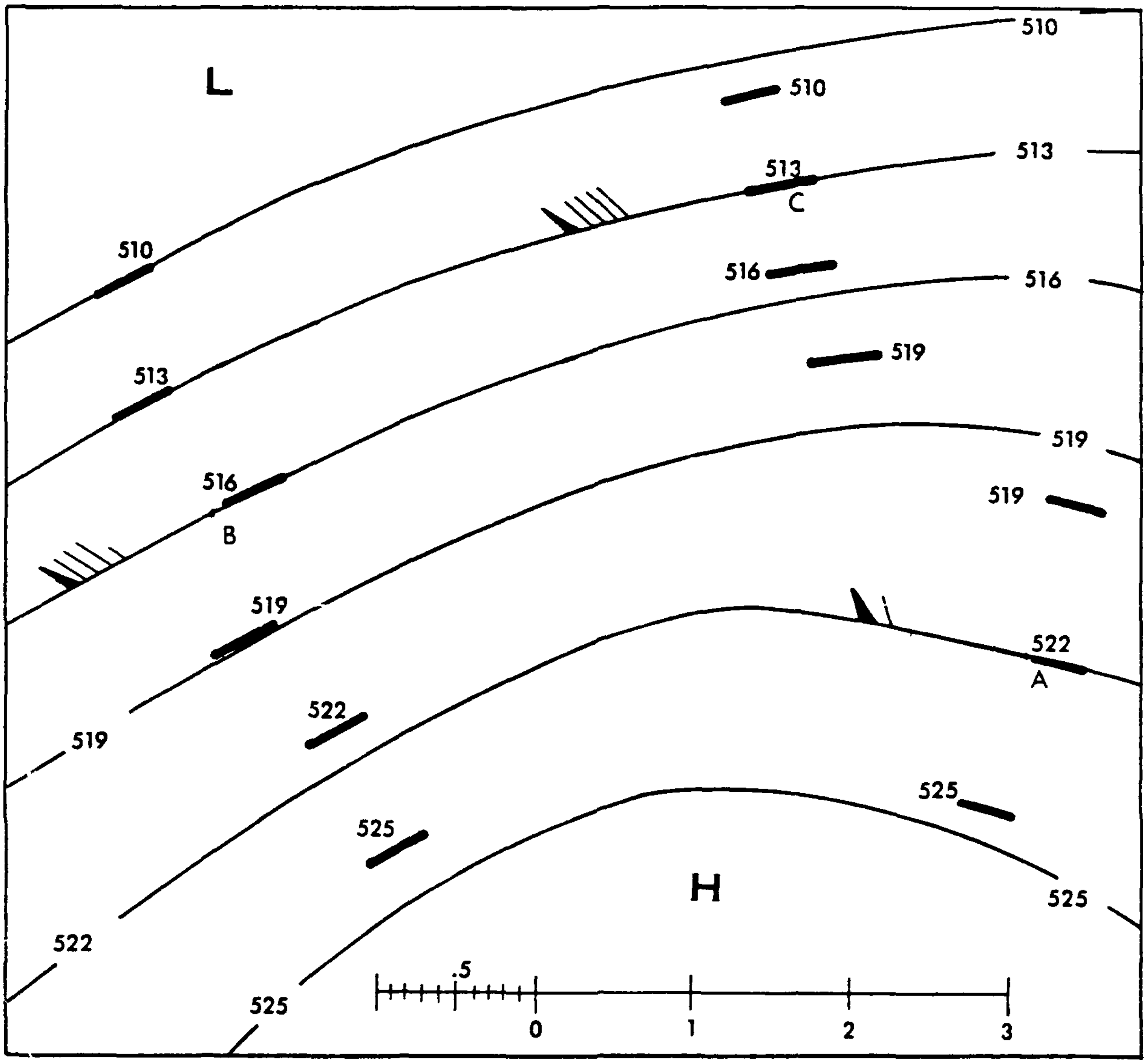


FIG. IV-B-2 500 mb chart with winds at different directions and speeds. The heavy segments show the geostrophic spacing at Stations A, B and C. With different wind directions at B and C there must be some anticyclonic curvature between them. Assuming that B was in an inflection point with almost straight flow the 513 contour was drawn through the spacing mark (513) and given anticyclonic curvature to pass through Station C. The 516 contour was drawn through B and given a larger spacing (south of the 516 mark at Station C). South of Station B some anticyclonic curvature was drawn so the spacing was increased. This procedure was continued to complete the map.

f. Prognosis. With a better analysis and knowledge of the convergence, vertical motion, vorticity, and moisture advection, the forecaster can make better prognostic maps and a better forecast than with only one station.

g. Forecast. A confident forecast may be given for the downwind stations. The timing for weather events that will enter the area is still a problem and the standard models are the best help.

3. Adding Non-synoptic Data. Data from aircraft or other sources (ships at sea, a tank crew) may become available. These data may not be at map time and are from almost any location or height. They can be very useful and must be used in continuity with time and adjusted on the synoptic map.

a. RECCO Report. These are the best and most complete airborne reports available. Plot all reports and start to adjust in time and space. TR 225 (Air Weather Service, 1960) has an excellent discussion on this subject.

b. Pilot Reports. Any information about tops and bases of clouds, cloud type, weather, turbulence, and aircraft icing helps to develop and extend the analysis. The pilot report needs to be adjusted in time and space the same as the RECCO data.

c. TARWI Reports. These are used in NATO for Target Weather Reporting. This code is not used as frequently as RECCO or PIREP so it is given in the Appendix, Part N. All these codes are given in AWSR 105-24 (Air Weather Service, 1983).

The information is primarily sky condition, visibiity (surface) and weather (surface). This information can be added to the single station data, or if successive reports are available, a single station procedure can be used for that target. Lack of wind data is the major handicap, but with a series of these data the forecaster can determine the synoptic condition. Thus, he can give the latest target weather report and a forecast for the next few hours.

The cloud and weather reports can be converted to a relative humidity report using the procedures in the Appendix, Parts C and D.

Consider how helpful the following TARWI report at D3-1800 LST in the southwest corner of the map would have been to the forecaster back in Chapter III-C-9c:

TARWI $YQL_aL_aL_aL_oL_oL_oG'G'$ 8343 SMOKE BLOWING SW.

He would have been certain that the front was past the corner of the map, as he drew it, and would have known about the low clouds to the rear of the front. Thus, he would have been more confident about his forecast for D4-0600 LST.

PART C AN EXAMPLE

1. Data. The observations and situation continues from Chapter III. The observations for D3-1800 LST to D4-0600 LST are given in Table IV-C-1.
2. Interpretation of the Data from D3-1800 LST to D4-0600 LST.
   a. Continuity. The front south of the station has become stationary as indicated on the map for D3-1800 LST, Fig. III-C-4.
   b. Clouds. By D3-2300 LST the forecaster should realize that overrunning is occurring because of the increasing amount and the lowering of the clouds.
   c. Radiation Cooling. The temperature holding steady indicates that the clouds are thick so little IR radiation is escaping, or warm air advection is occurring, or both.
   d. Wind Shear. The shear between the surface wind and the wind at the AS cloud level indicates almost no advection, but a strong wind shear. The gradient-level wind (GLW) will surely be from the ESE which would indicate weak warm air advection, and a low to the southwest.
   e. Pressure Falls. The continually falling pressure indicates that the low is moving toward the station, or intensifying, or both. The diurnal pressure change from 1800 to 2100 LST would indicate a rising pressure and almost no change until the next day at 0600 LST (Table II-C-4).
   f. D3-2355 Observation. The observation at D3-2355 appears to be incorrect, i.e., the sky condition does not seem to follow previous observations. Why has the ceiling lowered so fast? Should the forecaster assume a continuing lowering of the ceiling?
   g. Ceiling. By D4-0100 LST the ceiling has lifted to be in line with prior observations.
   h. Warm Front. The D4-0200 and D4-0300 LST observations give the meaning of the D3-2355 LST observation. The front probably is moving

TABLE IV-C-1 SURFACE OBSERVATIONS D3-1800 LST TO D4-0600 LST FOR STATION #3

DAY	TIME (LST)	SKY	VSBY	WW	SEALEVEL PRESSURE (MB)	SEALEVEL PRESSURE (IN)	TEMP (°F)	DEW POINT (°F)	RH	WIND DIR (°)	WIND SPEED (KT)	SKY DETAIL N_s	CC	$h_sh_sh_s$	N_s	CC	$h_sh_sh_s$	N_s	CC	$h_sh_sh_s$	REMARKS
D3	1855	120 SCT E300 OVC	7+		167	998	23	14		045	8	4	AS	120	10	CS					
D3	1954	E120 BKN 300 OVC	7+		163	997	24	14		045	11	6	AS	120	10	CS					
D3	2056	E120 BKN 300 OVC	7+		156	995	24	14		070	13	7	AS	120	10	CS					AS MVG ENE
D3	2157	E100 OVC	7+		160	996	24	15		070	9	10	AS	100							
D3	2255	E85 OVC	7+		160	996	24	16		045	12	10	AS	85							
D3	2355	E40 BKN 60 OVC	7+		160	996	2'	18		070	12	6	ST	40	10	ST	60				
D4	0052	E80 OVC	7+		146	992	25	18		070	7	10	AS	80							
D4	0155	E65 OVC	7+	S-	146	992	25	17		090	10	10	NS	65							SB 50
D4	.0253	W15 X	2	S-	142	991	25	20		070	12	10	FS	15							
D4	0337	15 .SCT E50 OVC	6	S-						070	12	4	FS	15	10	NS	50				
D4	0355	E50 OVC	7+	S-	121	985	25	21		070	13	10	NS	50							PRESFR
D4	0455	E45 OVC	7+		118	984	25	21		070	10	10	ST								SE 30
D4	0557	E16 OVC	7+	SG-	114	983	25	22		070	18	10	SC								SGB40

northward as a warm front (falling pressure, lowering ceilings and precipitation) and snow showers are forming ahead of the warm front. The low ceiling at D3-2355 occurred when low clouds associated with the developing snow showers moved past without precipitation.

i. Low Pressure. The continual falling pressure indicates that a major frontal wave is approaching, and the wind indicates it is to the southwest of the station.

3. Additional Data. Other observation teams also have been working and communications are established to the extent that data from two additional stations, #1 and #2, are available. The relative locations of the stations are shown in Fig. IV-C-1. The last 6 h of surface data for these stations are shown in Tables IV-C-2 and IV-C-3.

4. Interpretation of the Data from Station #1 D4-0000 to D4-0600 LST. Table IV-C-2.

   a. Air Mass. This station is in the cold air mass and the low is to the southwest of the station.

   b. Precipitation. The snow and rain are caused by overrunning of the warm front. The temperature is right for freezing rain.

   c. The Low. The small pressure change and winds indicate that the low is moving south of the station.

5. Evaluation of the Data from Station #2 D4-0000 LST to D4-0600 LST. (See Table IV-C-3.)

   a. Front. The warm front passed the station between 2358 and 0300 LST when the wind shifted. From 2358 to 0054 LST the temperature warmed 12° F and dewpoint increased 9° F.

   b. Low Center. The low pressure center is west of all stations

6. Upper Air Data. The rawinsonde data (Table IV-C-4) have been plotted as Figures IV-C-2, 3, and 4 for Stations #1, #2, and #3.

7. Analysis of the D4-0600 LST Data. The same single-station analysis procedures outlined in Chapter III should be utilized for each station before joining them together to make the analysis.

   a. Plotting the Maps. The basic data were plotted for each station then the GLW were added to the surface map. Because of the inversion at Station #3 (Fig. IV-C-4) the 2000 ft wind was selected to be the GLW. At Stations #1 and #2 the 3000 ft winds were selected for the GLW. The shear vectors computed from the GLW and the 700 mb level wind

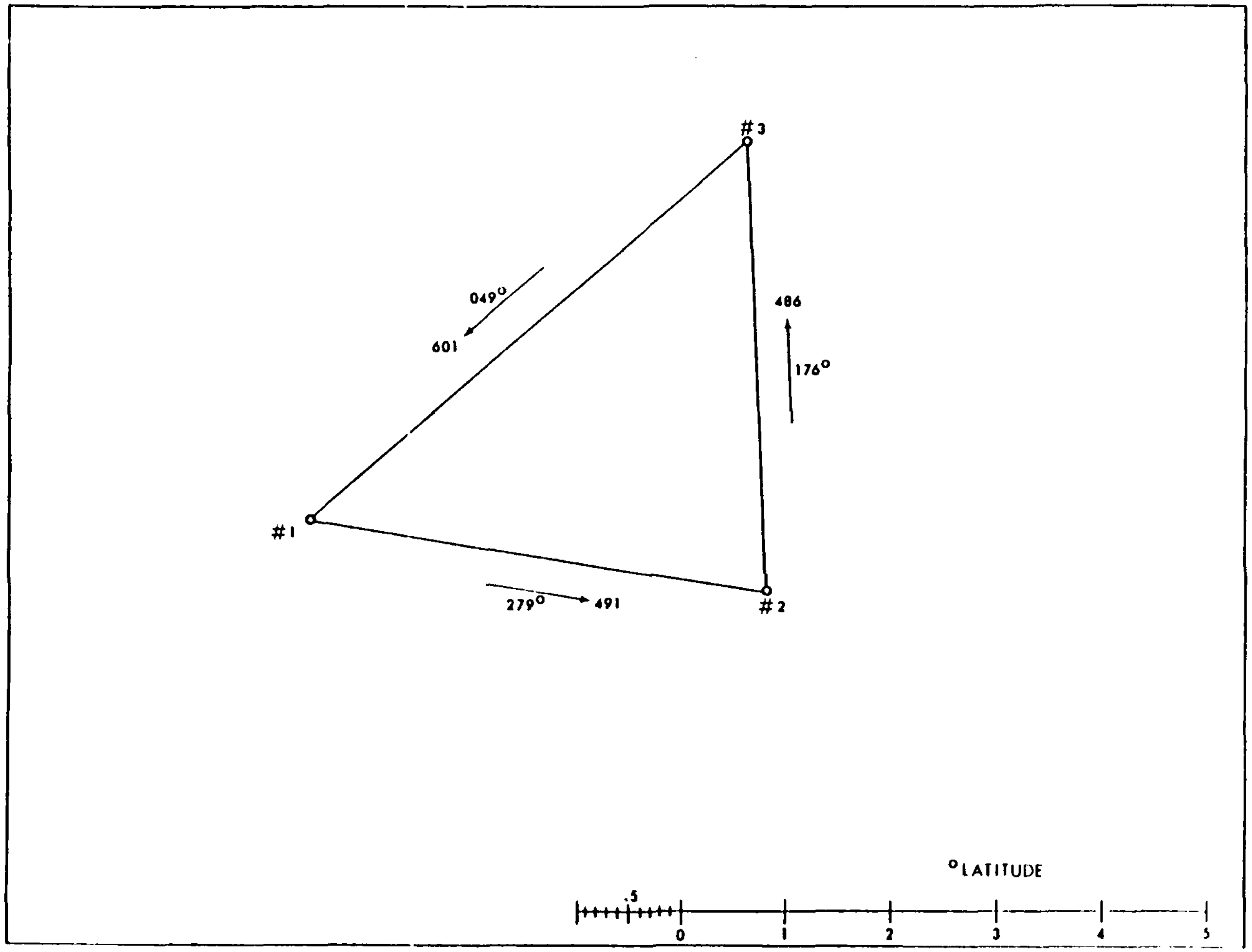


FIG. IV-C-1 Relative positions of the three stations with distances and directions (degrees from, as the wind is specified) of the sides of the triangle formed by the stations.

TABLE IV-C-2 SURFACE OBSERVATIONS D4-0000LST TO D4-0600 LST FOR STATION #1

DAY	TIME (LST)	SKY	VSBY	WW	SEA LEVEL PRESSURE (MB)	(IN)	TEMP (°F)	DEW POINT (°F)	RH	WIND DIR (°)	SPEED (KT)	SKY DETAIL N_s	CC	$h_sh_sh_s$	N_s	CC	$h_sh_sh_s$	N_s	CC	$h_sh_sh_s$	REMARKS
D3	2358	E6 OVC	7+		061	968	37	36		045	10	10	SC	6							SC MVG WSW 10
D4	0040	E3 OVC	5	F						045	13	10	ST	3							
D4	0058	E3 OVC	5	F	058	967	36	36		045	12	10	ST	3							
D4	0125	E7 OVC	7+			965				025	14	10	ST	7							
D4	0156	E6 OVC	7+		051	965	35	34		025	13	10	ST	6							
D4	0258	E8 OVC	7+		047	964	33	31		025	15	10	ST	8							
D4	0355	E7 OVC	8	S-	041	962	32	31		045	14	10	ST	7							SB 38
D4	0458	E9 OVC	8		027	958	33	32		025	14	10	ST	9							SE 15
D4	0555	E7 OVC	8	R-	024	957	33	32		045	12	10	ST	7							RB18

TABLE IV-C-3 SURFACE OBSERVATIONS D4-0000 LST TO D4-0600 LST FOR STATION #2.

DAY	TIME (LST)	SKY	VSBY	WW	SEA LEVEL PRESSURE (MB)	(IN)	TEMP (°F)	DEW POINT (°F)	RH	WIND DIR (°)	SPEED (KT)	SKY DETAIL N_s	CC	$h_sh_sh_s$	N_s	CC	$h_sh_sh_s$	N_s	CC	$h_sh_sh_s$	REMARKS
D3	2358	E20 OVC	4	F	079	975	48	47		045	5	10	ST	20							ST MVG ENE
D4	0030	E20 OVC	6	F		975				205	16	10	ST	17							
D4	0054	E17 OVC	7		076	974	60	56		205	14	10	ST	17							
D4	0157	E13 OVC	7		076	974	61	57		180	12	10	ST	13							
D4	0258	E12 BKN 32 OVC	8		076	974	62	58		180	12	7	ST	12	10	SC	32				SC LYR MVG NE RPDY
D4	0356	E14 BKN 32 OVC	10		071	973	62	58		180	16	8	ST	14	10	SC	32				
D4	0458	E16 OVC	10		075	974	62	58		180	12	10	ST	16							
D4	0555	E16 OVC	10		071	973	63	58		180	15	10	ST	16							

TABLE IV-C-4

Upper air data for stations for D4-0600 LST.

	#1	#2	#3
Sfc	040/12	180/15	060/18
1000 ft	---	190/26	080/17
2000 ft	090/15	200/35	100/25*
3000 ft	160/19*	220/40*	110/24
4000 ft	190/30	220/43	110/20
5000 ft	210/44	225/47	140/20
6000 ft	220/44	225/50	180/23
7000 ft	230/45	225/50	220/32
8000 ft	230/53	225/51	220/43
9000 ft	230/58	225/54	220/48
10000 ft	230/63	225/55	230/47
12000 ft	230/71	235/70	235/56
16000 ft	235/86	235/90	240/71
20000 ft	235/98	240/103	250/77
850 mb	200/36	225/48	120/20
700 mb	230/62	225/55	230/48
500 mb	235/95	240/100	240/78
300 mb			250/103
850 H	353	433	390
850 T	8.8	10.2	1.2
850 $T-T_d$	0.6	2.7	1.3
700 H	941	021	941
700 T	2.0	2.5	-4.2
700 $T-T_d$	5.0	8.0	0.6
500 H	555	566	552
500 T	-17.5	-16.0	-18.8
500 $T-T_d$	x	8.0	2.4

*Gradient-level Wind

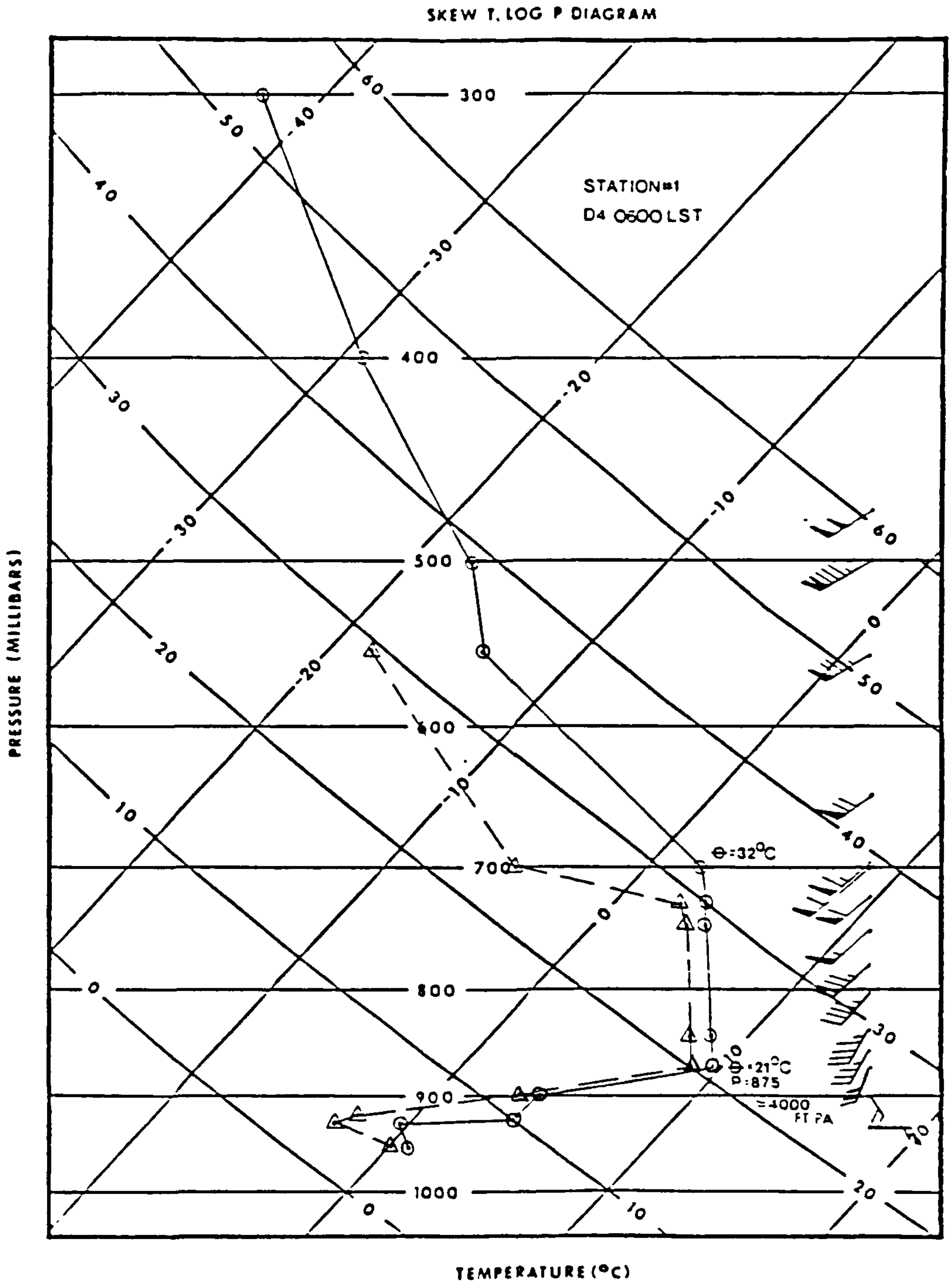


FIG. IV-C-2 Rawinsonde data for Station #1 for D4-0600 LST. Front is at p = 875 mb with a potential temperature value of 21° C. The break of lapse rate at p = 700 mb and θ = 32° C is probably subsidence.

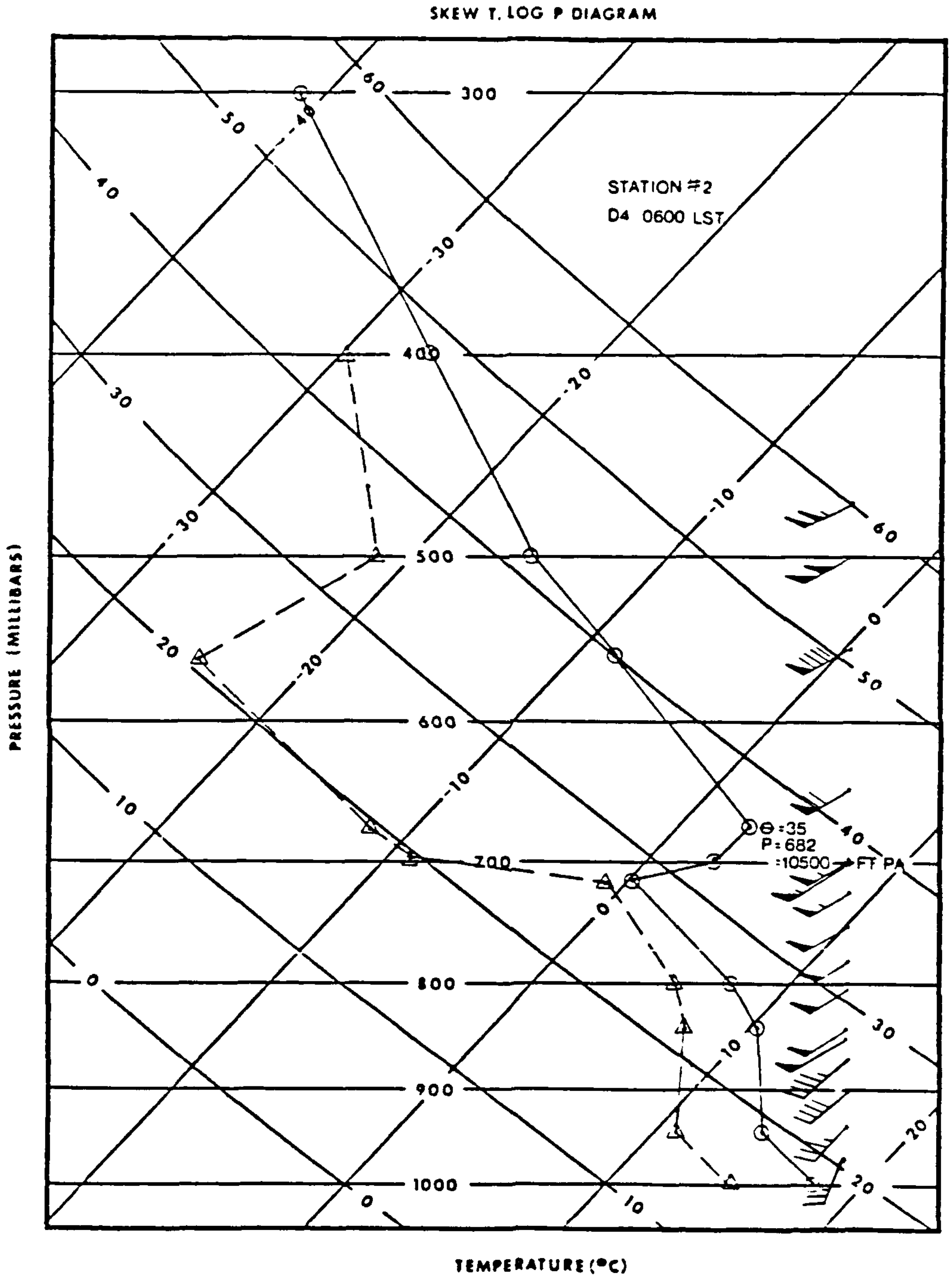


FIG. IV-C-3 Rawinsonde data for Station #2 for D4-0600 LST. The front is not over this station but the subsidence layer with θ = 32° C is same layer at 32° C at Station #1.

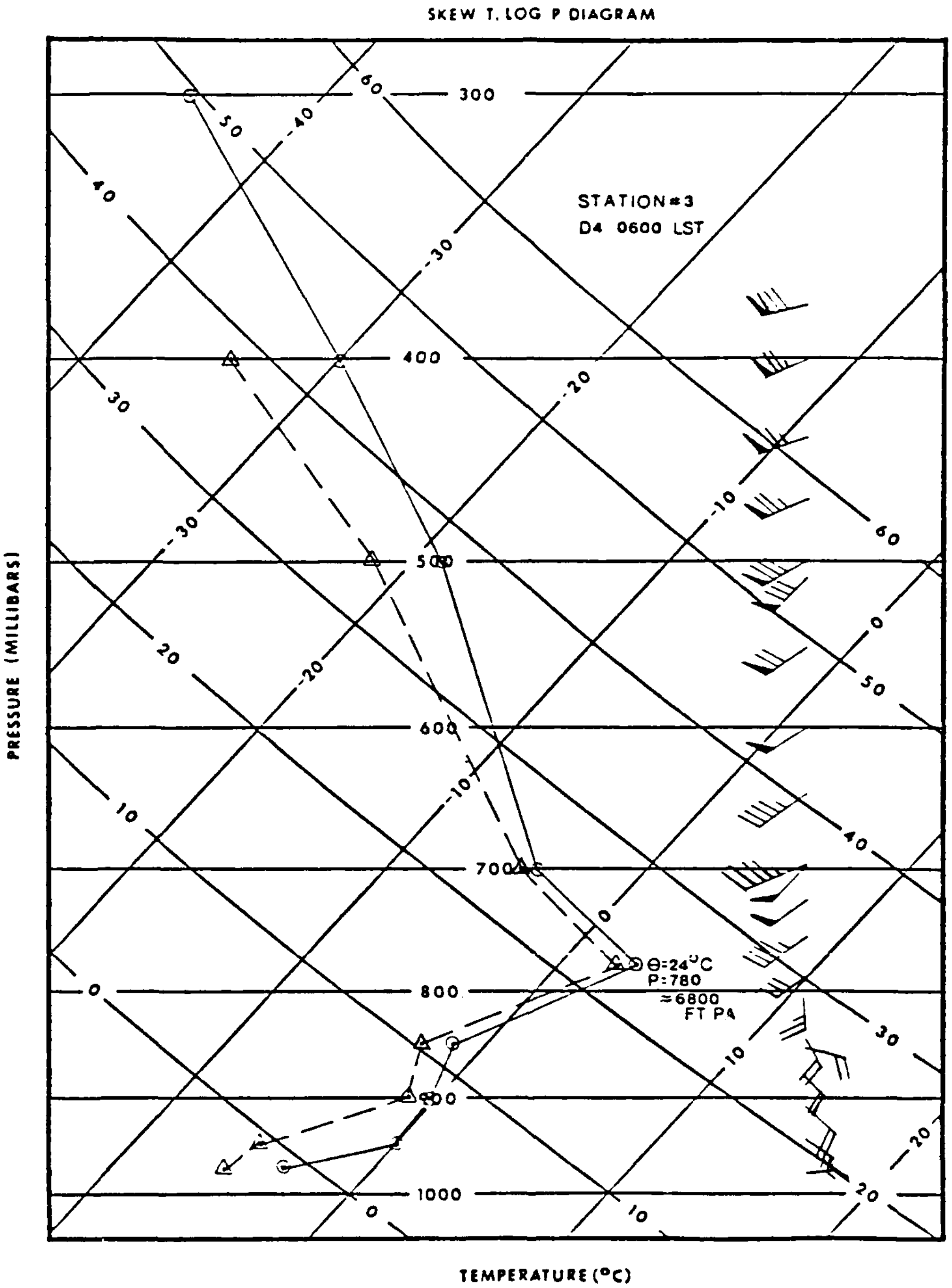


FIG. IV-C-4 Rawinsonde data for Station #3 for D4-0600 LST. Front is at p = 780 mb and θ = 24° C. Compare with θ of 21° C at Station #1.

were computed and plotted on the surface map. The shear vectors also were plotted on the 850, 700, and 500 mb maps as described in Chapter III, along with the data for each level. (See Figs. IV-C-5, 7, 9, and 12.)

b. Station #2 D4-0600 LST. The warm front passed #2 soon after D4-0000 LST. The GLW is from the SSW and surface wind in the cold air is from 045°, so the front will not move fast - maybe at 5 kt. Therefore, it will displace about 0.5° Lat in the 6 h period to 0600 LST so it should be placed north of Station #2 and parallel to the shear vector of the GLW and the 700 mb wind.

c. Station #1 D4-0600 LST. The sounding (see Fig. IV-C-2) shows the top of the inversion at 875 mb, and having a θ of 21° C which would convert to a temperature of 7.5° C at 850 mb. The 875 mb level is about 4000 ft <u>pressure altitude</u> (PA) and the surface is 1300 ft MSL. Warm air advection is indicated so the front is a warm front. Assuming a slope of 1:200 for the warm front, the front is 1.5° Lat from Station #1 at the surface

$$\frac{4000-1300}{5280} \times \frac{200}{69} = 1.5° \text{ Lat}$$

and

$$\frac{5000-4000}{5280} \times \frac{200}{69} = 0.5° \text{ Lat}$$

from the station at the 850 mb level. The orientation of the front is determined by the direction of the GLW-700 mb wind shear vector. These distances could be computed using the actual height of the 850 mb level, but the error in estimating the slope does not require such close calculations, so the ease of reading the pressure altitude is utilized.

d. Station #3 D4-0600 LST. The sounding (see Fig. IV-C-4) shows the top of the front at 780 mb (7000 ft <u>PA</u>) with a θ of 24° C (10° C at 850 mb). The station is 800 ft MSL. Using a slope of 1:200 the surface position would be

$$\frac{7000-800}{5280} \times \frac{200}{69} = 3.4° \text{ Lat}$$

from Station #3. It is orientated parallel to the GLW-700 mb wind shear vector. The front is

$$\frac{7000-5000}{5280} \times \frac{200}{69} = 1.1° \text{ Lat}$$

from Station #3 at the 850 mb level. At the 700 mb level the front would be

$$\frac{9600-7000}{5280} \times \frac{200}{69} = 1.4° \text{ Lat}$$

distant from Station #3. At the 700 mb height the front should be a little flatter so 1.5° or 1.6° Lat could be used. The isotherm along the front is -5° C. The value of -5° C at 700 mb was determined by following the θ value to the 700 mb level. Note how well the potential temperature at the top of the front was conserved during the last 12 h.

e. Placing Front. When 3.4° Lat from Station #3 is located along a line normal to the GLW-700 mb wind shear vector, the position is very close to the location estimated from Station #2 in part 5a. With the location of the two points just north of Station #2, the distance south of Station #1 (computed in part 7c) and the direction of the shear vectors, the front may be drawn on the surface map. (See Fig. IV-C-5.)

f. Surface Map. The cold front on the western edge of the map was drawn to complete the model and remind the forecaster that a low with frontal wave is expected.

Using the geostrophic wind tables (Appendix, Part I) the spacing of the isobars may be marked (shown by the heavy lines). All the determined spacings fit except at Station #1. Because of the cyclonic curvature the isobars were drawn with closer spacing then would be indicated by the geostrophic wind.

The surface weather and nephanalysis are shown in Fig. IV-C-6 for D4-0600 LST. The snow showers at Station #3 were moved northeastward from the station. The rain area at Station #1 extended east and west of the station with snow to the north. Showers were placed along and ahead of the front because of the instability shown on the stability chart Fig. IV-C-14 which is discussed later. The nephanalysis was made using the present and past ceilings of the three stations.

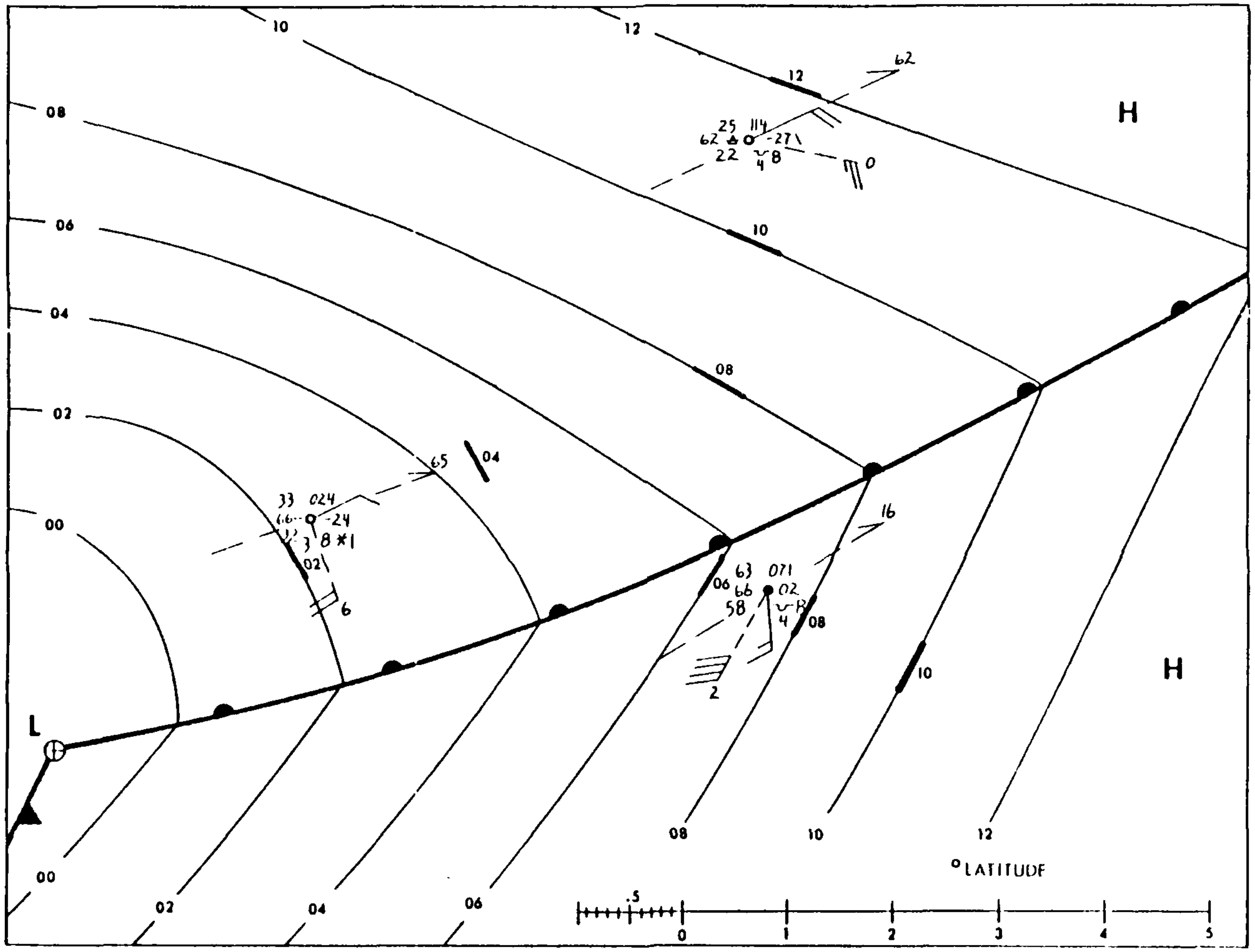


FIG. IV-C-5 Surface map for D4-0600 LST. Heavy lines on the isobars indicate spacing to match the gradient wind speed. Dashed winds are gradient-level winds and the arrow is the shear vector from the gradient-level wind and the 700 mb wind.

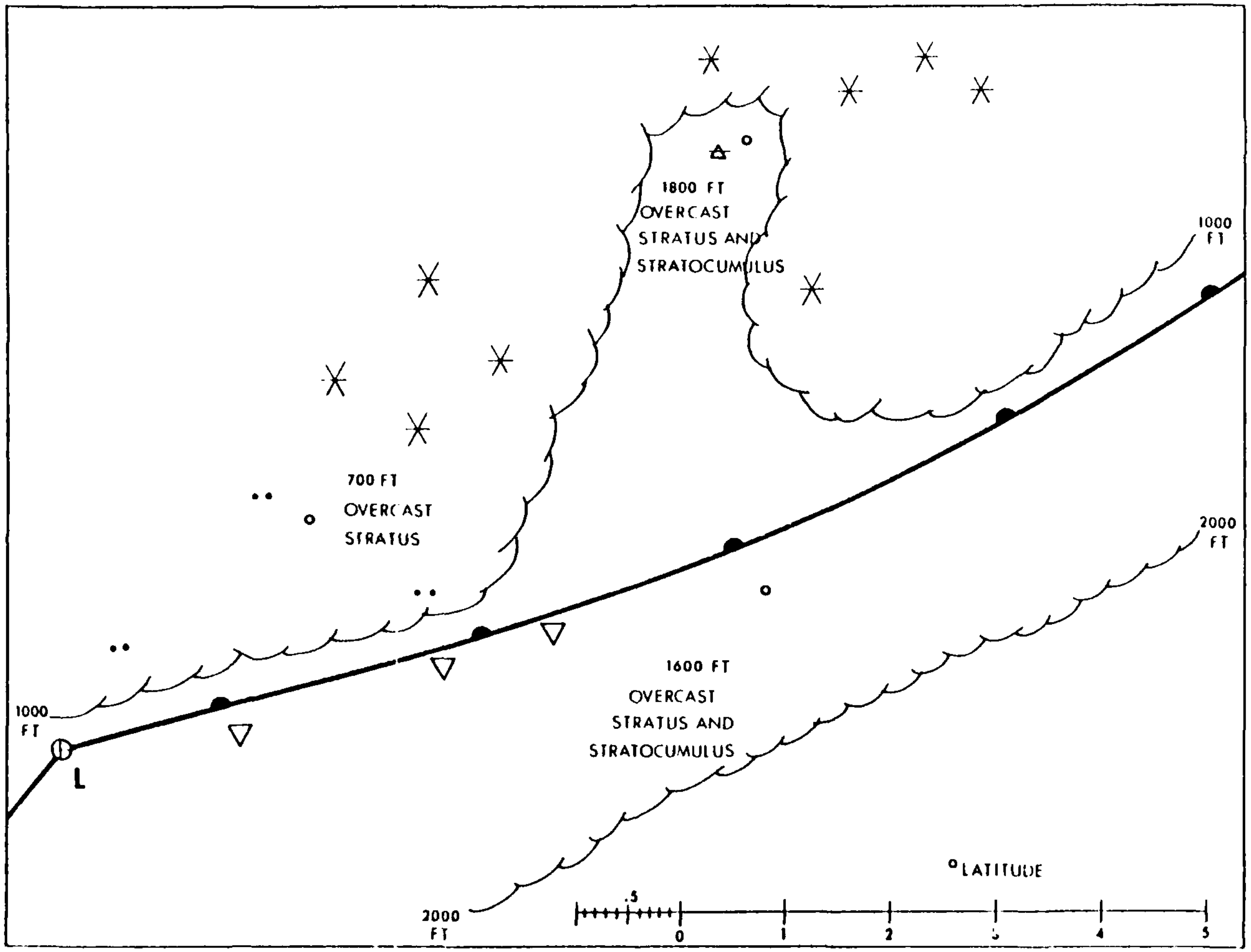


FIG. IV-C-6 Surface weather analysis and nephanalysis for D4-0600 LST.

g. 850 mb Map. The 850 mb map (Fig. IV-C-7) is drawn to show the front at the distances from the stations calculated in 7c and 7d and properly stacked with the surface. The 30 m contours are spaced using the geostrophic wind table in Appendix, Part J. The spacing is indicated by the heavy section of the contours. The isotherms are at 2° C intervals and the 6° C isotherm is placed just north of the front and the others in relation to the temperature gradient read from the Appendix, Part L. A check of the plotted RAOB data shows the large temperature difference between the two air masses which indicates a close packing as was also indicated by the shear vector.

h. The 850 mb Moisture Chart. Fig. IV-C-8 is an analysis of the 850 mb T_d and 850 mb mixing ratio (r) (solid lines labeled with dew point temperature at the left end of the line and mixing ratio on the right) and $T-T_d$ (dashed). This chart presents the moisture pattern of a warm front with precipitation caused by overrunning. Thus the isodrosotherms are close to their saturation isotherm on the northern two-thirds of the map. Only on the southern edge of the map might be a cloud free area at the 850 mb level. In practice, the moisture may be placed on the 850 mb chart by using color (green) but was kept separate for clarity in this presentation.

i. The 700 mb Chart. The data were plotted on the 700 mb chart (Fig. IV-C-9) and the spacing of the contours are marked (heavy mark). When the contours are drawn to fit the spacing a trough is implied just west of Stations #2 and #3. On the surface and 850 mb charts a trough is on the western edge of the map. With warm air advection occurring at both Station #2 and #3 at the 700 mb level and warm air advection from the surface to 700 mb (see GLW-700 thickness vector on surface map) this trough in Fig. IV-C-9 does not fit the conditions and the contour analysis must be incorrect. It needs a curvature adjustment as described in Fig. IV-B-1c. The adjustment is made by introducing anticyclonic curvature at Station #2 and spreading the gradient as shown in Fig. IV-C-10. The 294 contour is drawn with only slight curvature from Station #1 to #3. To the southeast anticyclonic curvature is shown in the contours and a wider spacing than the geostrophic spacing indicates. To the northwest of the 294 contour the contours were drawn with cyclonic curvature, because the

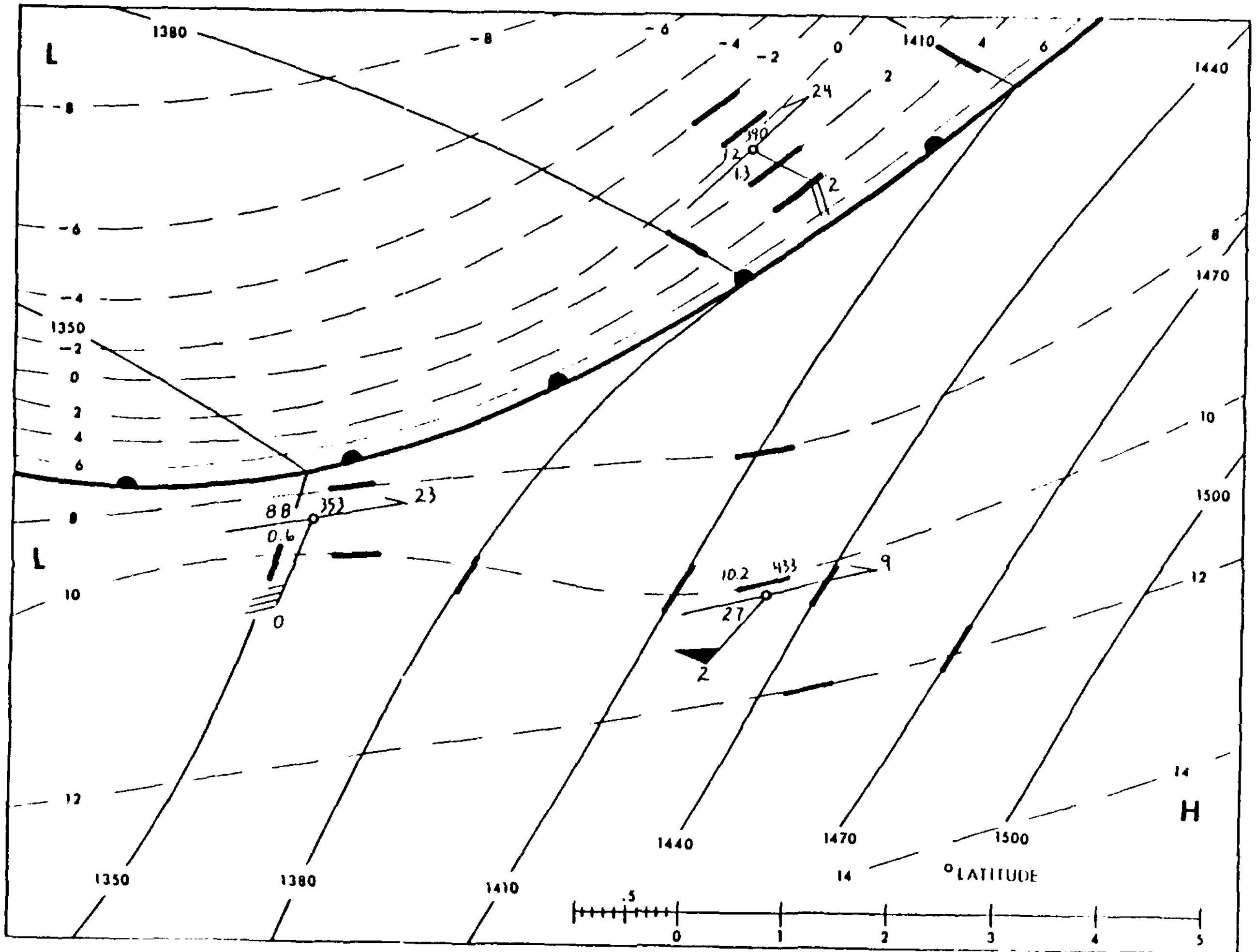


FIG. IV-C-7 850 mb analysis for D4-0600 LST. Heavy lines on contours and isotherms (dashed) are the geostrophic spacing. Arrows are the 4000 ft to 6000 ft wind shear vectors.

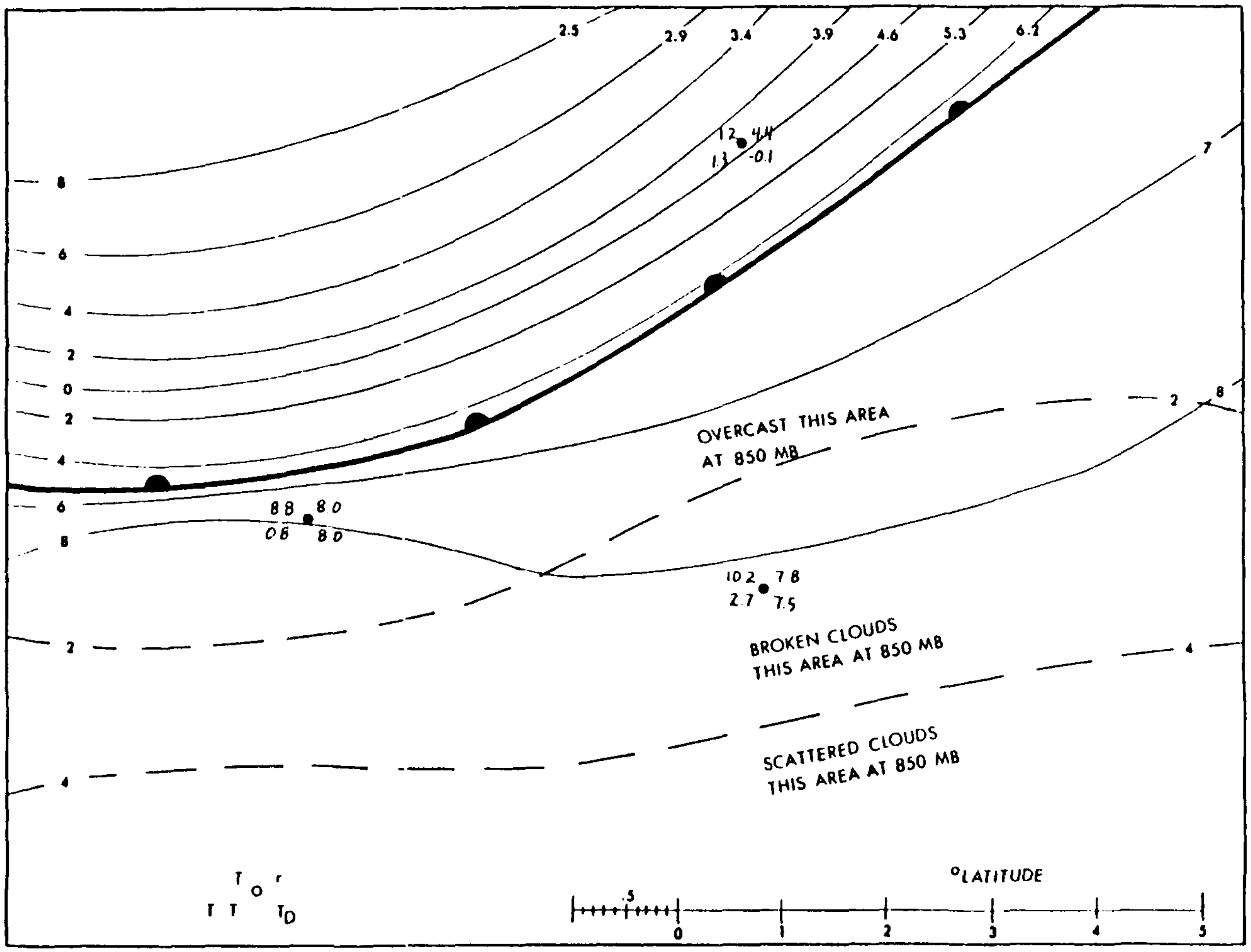


FIG. IV-C-8 Analysis of 850 mb isodrosotherms (marked as dew point temperature ° C) on the left and mixing ratio (r) on the right. The $T-T_d$ are dashed lines on the chart. The plotting model is in the lower left hand corner.

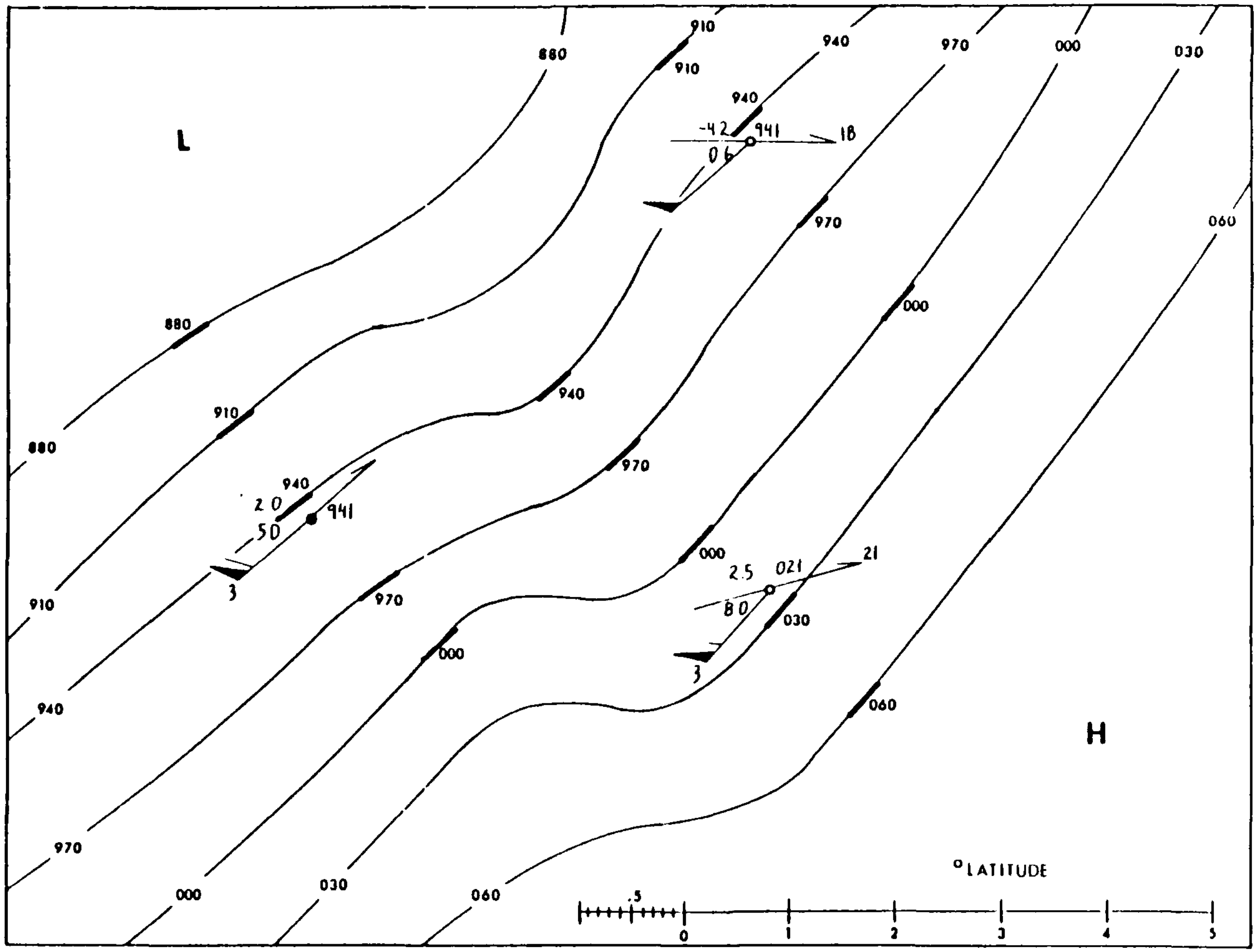


FIG. IV-C-9 An analysis of the 700 mb chart for D4-0600 LST using the gradient spacing for the winds. This is an incorrect analysis.

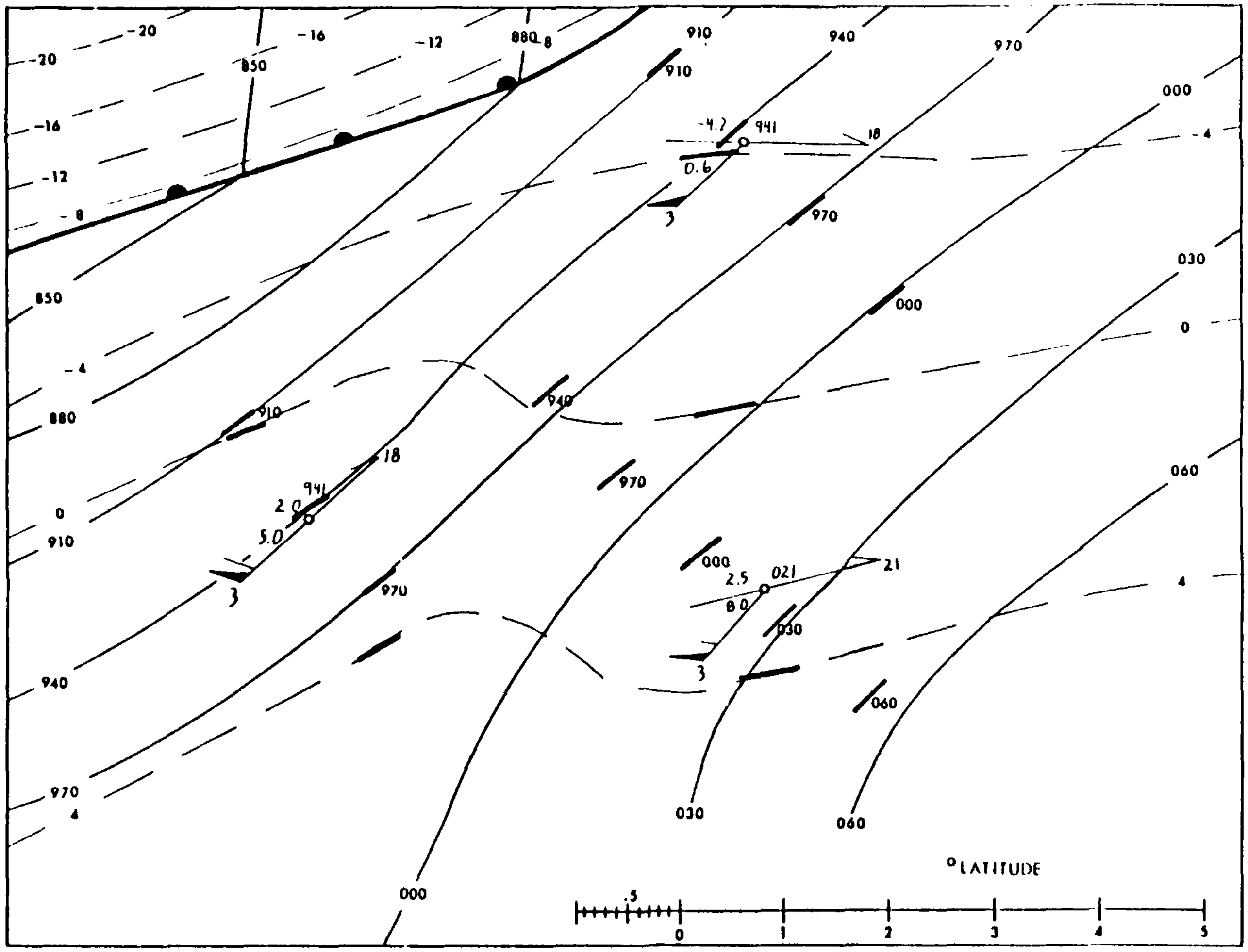


FIG. IV-C-10 700 mb analysis for D4-0600 LST. Heavy lines on the map indicate geostrophic spacing of both contours and isotherms (dashed). The arrows are the shear vector from the wind at 2000 ft. The requirement for anticyclonic curvature expands spacing of lines.

front is in the area. The potential temperature at the top of the frontal zone on the RAOB (Fig. IV-C-4) would indicate a -5° C isotherm along the front so the -8° C is placed in the cold air, close to and parallel to the front at 700 mb with a strong thermal gradient to the northwest.

j. 700 mb Moisture Chart. At both Stations #1 and #2 the 700 mb level is much drier than the layers just below as shown by the radiosonde data, Fig. IV-C-2 and 3. The moist overrunning over the front extends well above the 700 mb level over Station #3, however, because the cold air is deeper. The $T-T_d$ = 2°, 4°, and 6° C are shown in Fig. IV-C-11. As in the case of the 850 mb chart the moisture could be added to the contour analysis by using color.

k. 500 mb Chart. The same type of corrections to adjust for the geostrophic wind gradient that were applied to the 700 mb chart had to be used on the 500 mb chart. The geostrophic spacing is marked and the amount of deviation because of the anticyclonic curvature can be seen on Fig. IV-C-12. The weak thermal pattern shown by the dashed lines would be expected in the warm sector of a wave cyclone.

At Station #3 the 12 h height change was +20 m. A study of the z-t chart (Fig. IV-C-13) shows that Station #3 is still in the ridge and should have anticyclonic curvature. The trough is west of Station #1 which allows proper hydrostatic stacking.

l. The Instability Chart. The shear of shears vector (instability shear) for each station and the stability indices were computed for each station and plotted on the map. The location of the surface front was traced on the chart, Fig. IV-C-14. The TTI is presented on the analysis. The area of instability along the front is indicated with the shear of shears vectors at Station #3 from the east and Station #2 from the west. Station #1 has unstable air over it and more unstable air to the southwest.

8. <u>Evaluation of the Vertical Motion and Vorticity</u>.

a. Calculation of Vertical Motion. The vertical motion at 500 mb was calculated using both (IV-4) and (IV-11) for comparison. These calculations will be referred to as method 1 and method 2, respectively. In both cases, the calculations proceeded by first calcula-

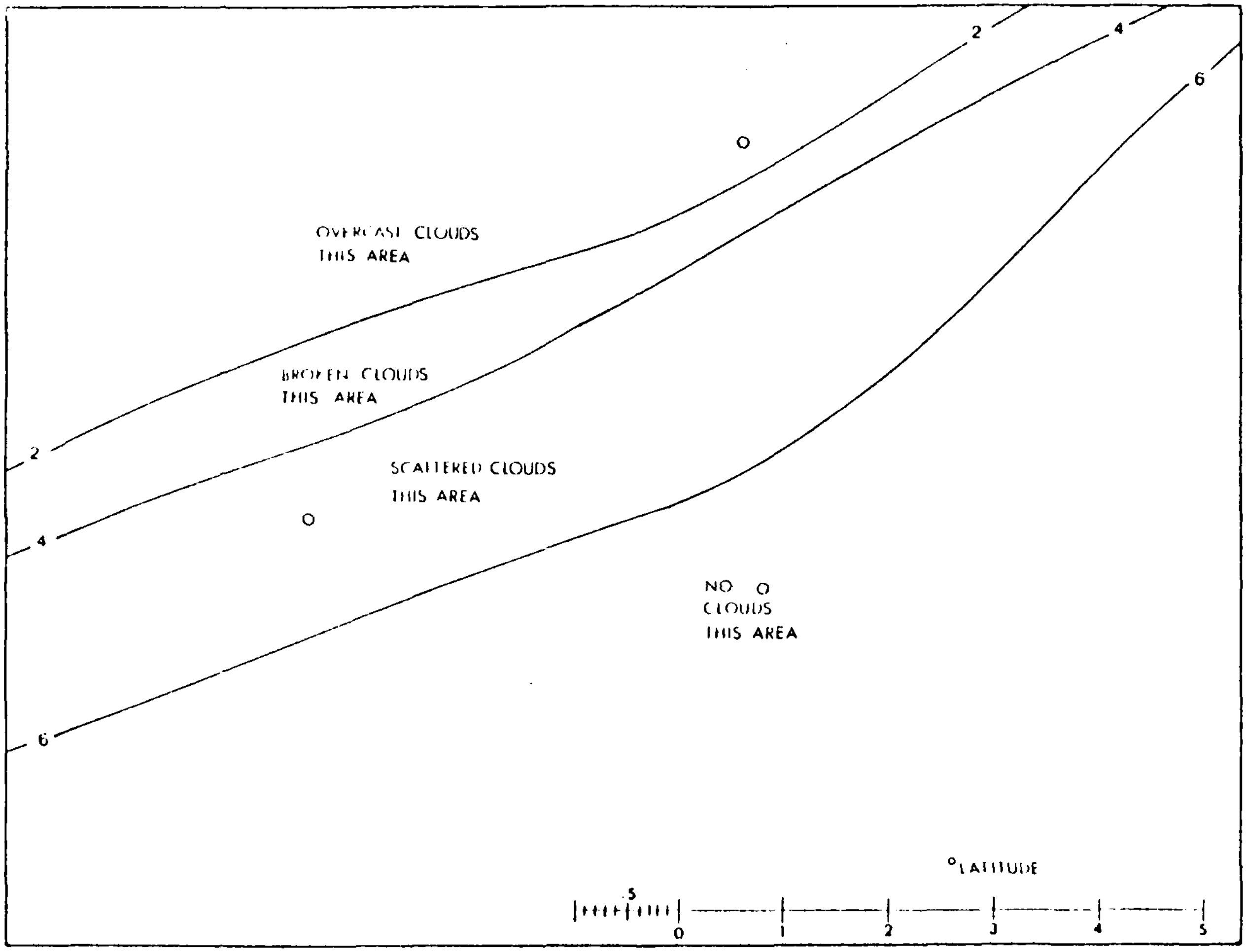


FIG. IV-C-11 The moisture analysis of the 700 mb chart for D4-0600 LST. $T-T_d$ in ° C are used and the cloud amounts expected are shown.

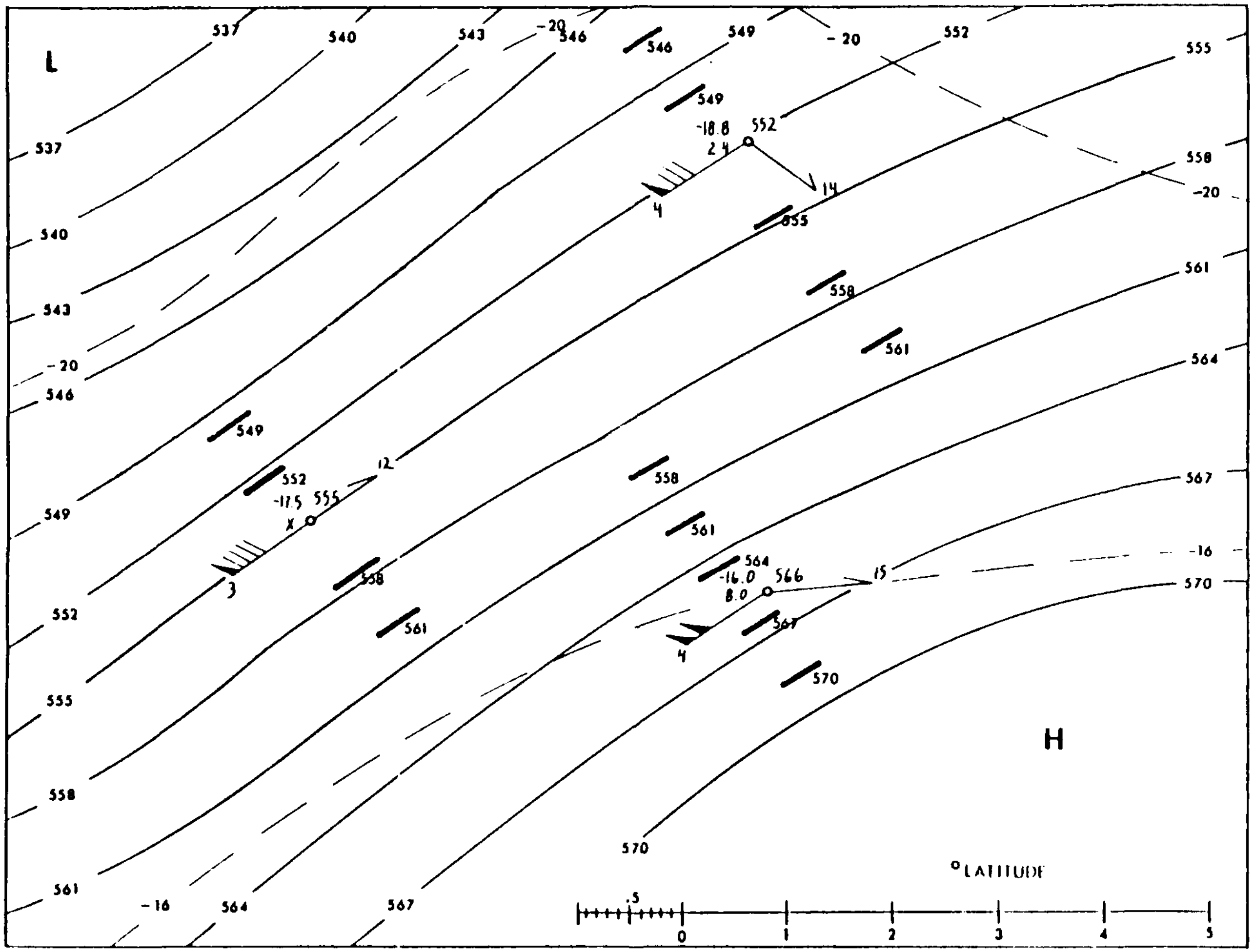


FIG. IV-C-12 500 mb analysis. The short heavy lines indicate the gradient spacing for the contours and isotherms (dashed). The amount of anticyclonic curvature to make the lines fit can be seen. Arrows are the shear vector between the winds at 16000 and 20000 ft.

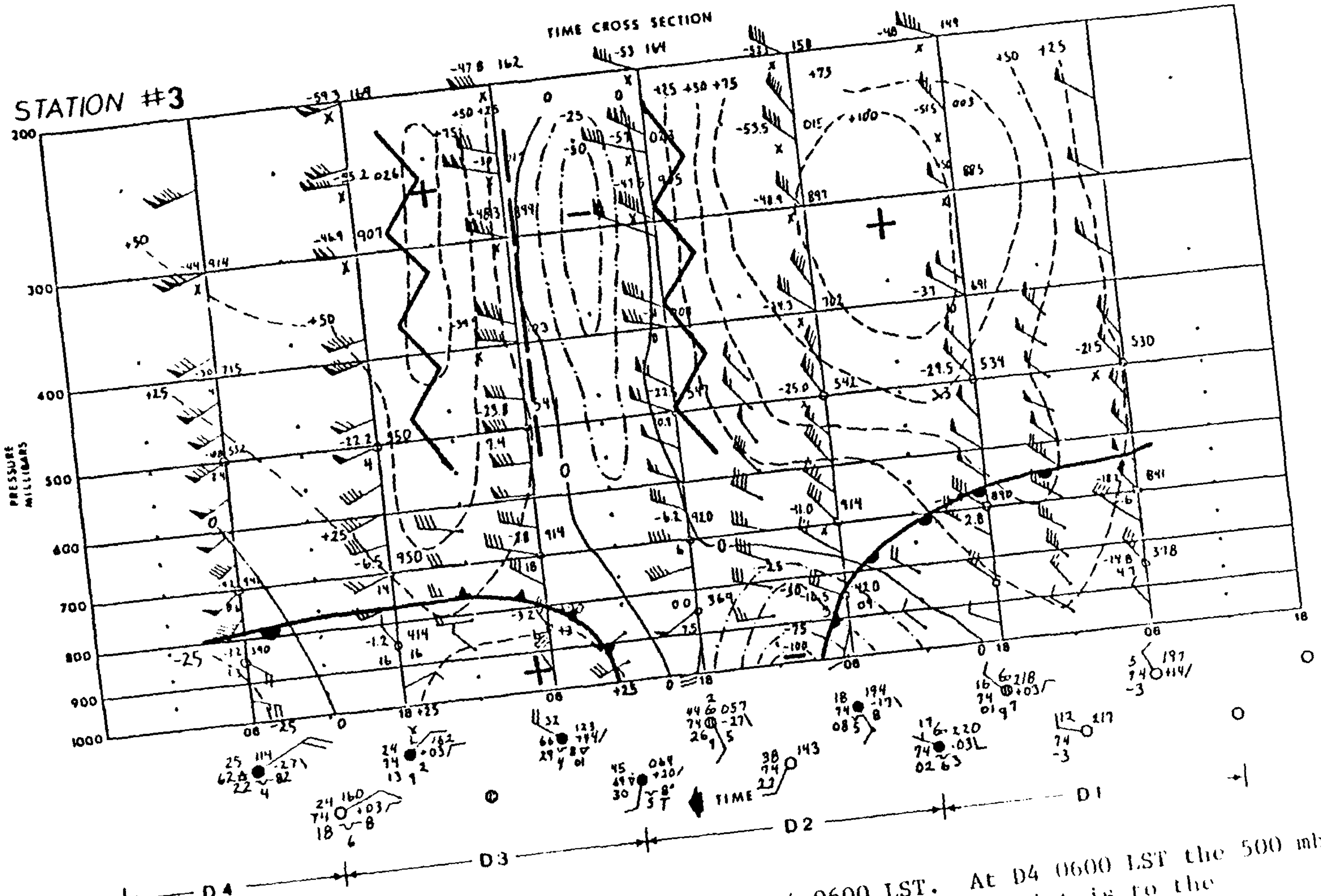


FIG. IV-C-13 The z-t chart continuity up to D4-0600 LST. At D4 0600 LST the 500 mb level is still in the ridge and the inflection point is to the southwest.

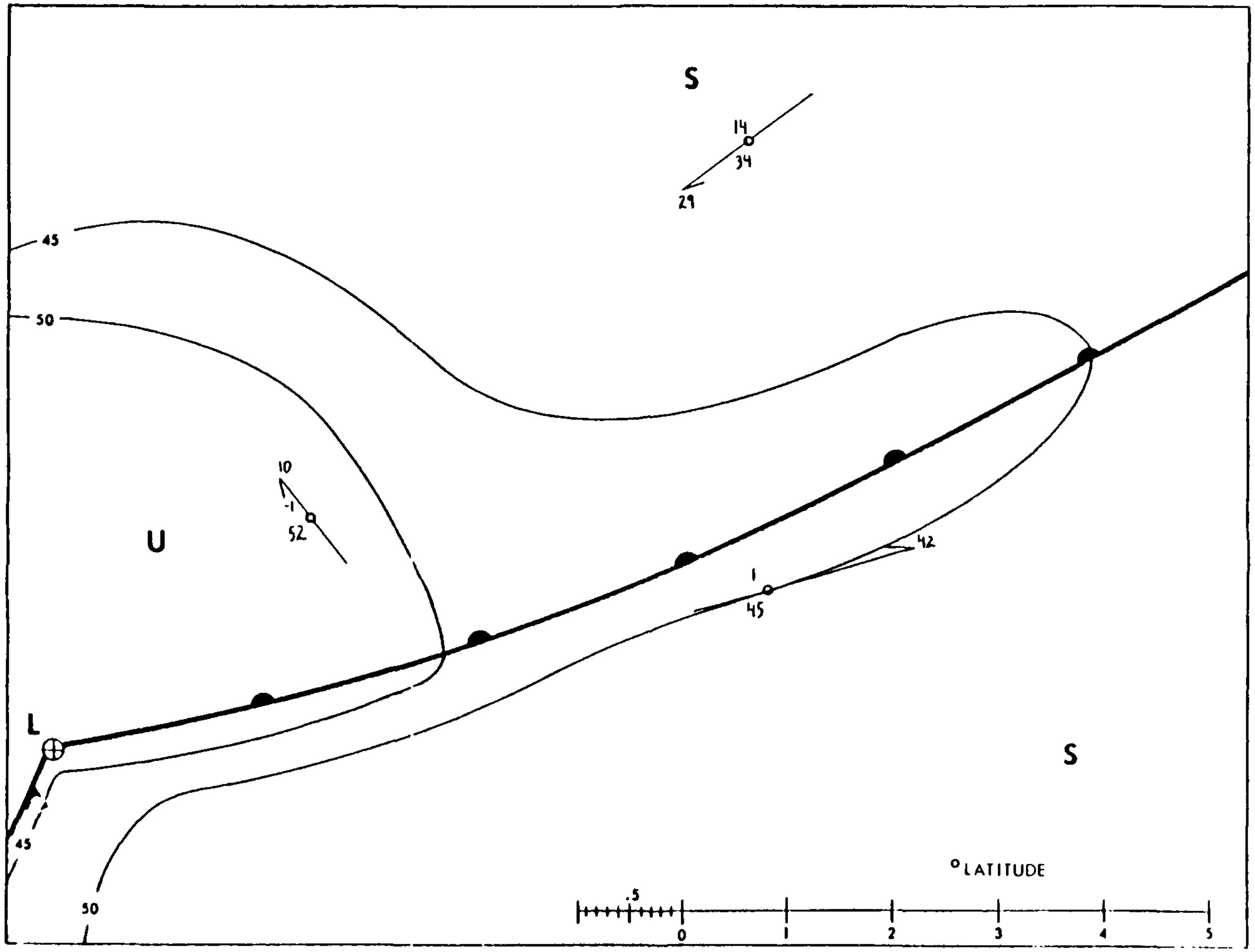


FIG. IV-C-14 Instability chart for D4-0600 LST. The Shear of Shears for each station is plotted and the Showalter Index is plotted above the station with the Total-Totals Index below the station. The isolines are TTI. The front was added to show the relationship between the front and the unstable air.

ting w at 2000 ft, then at 4000 ft, and finally at 500 mb (actually 5580 m).

The wind observations needed for the calculations appear in Table IV-C-4. From these, the velocity components at the three stations can be calculated using (IV-3). These values and the other pertinent information for the three stations are given in Table IV-C-5. Table IV-C-6 provides the remaining information required for the three sides of the triangle (see Fig. IV-C-1). The area enclosed by the triangle in this case is $A' = 112,315 \times 10^6 \, m^2$. It will be recalled that method 1 uses (IV-4), which is repeated here for the convenience of the reader.

$$\begin{aligned} \overline{w}_T = \overline{w}_B \; - \frac{1}{A'}[& (\overline{v}_a \sin\beta_a - \overline{u}_a \cos\beta_a)A_a \\ & + (\overline{v}_b \sin\beta_b - \overline{u}_b \cos\beta_b)A_b \\ & + (\overline{v}_c \sin\beta_c - \overline{u}_c \cos\beta_c)A_c] \; . \end{aligned} \tag{IV-4}$$

Thus, w by method 1 at 2000 ft is given by

$$\begin{aligned} w_{2000} = - \frac{1}{112,315 \text{ X } 10^6} & [(498 \sin 279° + 138 \cos 279°)132,373.6 \text{ X } 10^3 \\ & + (556 \sin 176° + 363 \cos 176°)175,312.8 \text{ X } 10^3 \\ & + (-178 \sin 49° + 809 \cos 49°)216,119.6 \text{ X } 10^3] \\ = 0.3 \text{ cm s}^{-1}. \end{aligned}$$

The same procedure leads to $w_{4000 \text{ ft}} = 0.9$ cm s^{-1} and $w_{500} = 3.1$ cm s^{-1}.

The quantity Δ (IV-7), which is required for the computation using method 2, has the value, $\Delta = 223,356 \text{ km}^2$. The average terrain height at the centroid of the triangle is 275 m. The values of u_o and v_o, calculated from (IV-8 a,b) for the surface centroid winds are given in Table IV-C-7. Also given in Table IV-C-7 are the components of the terrain gradient $(\partial z/\partial x)o$ and $(\partial z/\partial y)o$, calculated from (IV-10 a,b). When these values are used in (IV-9), an orography-induced vertical motion at the centroid of 0.14 cm s^{-1} is obtained. The various derivatives of the velocity components required in this method were calculated from (IV-8, c-f), using the

TABLE IV-C-5

Station parameters and data

computed from rawinsonde data in Table IV-C-4.

	#1	#2	#3
x	- 330 km	151.8 km	123.3 km
y	- 93 km	- 167.6 km	300.4 km
Height (MSL)	391 m	78.3 m	237.1 m
u (sfc)	- 397 cm s^{-1}	0	- 802 cm s^{-1}
v (sfc)	- 473 cm s^{-1}	772 cm s^{-1}	- 463 cm s^{-1}
u (2000 ft)	- 772 cm s^{-1}	616 cm s^{-1}	-1266 cm s^{-1}
v (2000 ft)	0	1692 cm s^{-1}	233 cm s^{-1}
u (4000 ft)	268 cm s^{-1}	1422 cm s^{-1}	- 967 cm s^{-1}
v (4000 ft)	1520 cm s^{-1}	1694 cm s^{-1}	352 cm s^{-1}
u (500 mb)	4003 cm s^{-1}	4455 cm s^{-1}	3475 cm s^{-1}
v (500 mb)	2872 cm s^{-1}	2572 cm s^{-1}	2006 cm s^{-1}

TABLE IV-C-6

Parameters and data for triangle sides used in methods 1 and 2 computations.

	side a	side b	side c
Average surface base height (MSL)	340 m	235 m	250 m
Length	491 km	468 km	601 km
β	279°	176°	49°
Surface to 2000 ft (300 m)			
$\bar{u}$	-138 cm s^{-1}	-363 cm s^{-1}	-809 cm s^{-1}
$\bar{v}$	498 cm s^{-1}	556 cm s^{-1}	-178 cm s^{-1}
Area	132,373.6 X 10^3 m^2	175,312.8 X 10^3 m^2	216,119.6 X 10^3 m^2
2000 ft to 4000 ft (600 m)			
$\bar{u}$	383 cm s^{-1}	-49 cm s^{-1}	-684 cm s^{-1}
$\bar{v}$	1226 cm s^{-1}	990 cm s^{-1}	524 cm s^{-1}
Area	29,931.36 X 10^4 m^2	28,529.28 X 10^4 m^2	36,636.96 X 10^4 m^2
4000 ft to 5580 m			
$\bar{u}$	2654 cm s^{-1}	2214 cm s^{-1}	1695 cm s^{-1}
$\bar{v}$	2138 cm s^{-1}	1629 cm s^{-1}	1688 cm s^{-1}
Area	2,140.76 X 10^6 m^2	2,040.48 X 10^6 m^2	2,060.36 X 10^6 m^2
	$x_1y_2-x_2y_1=69{,}425$ km^2	$x_2y_3-x_3y_2=66{,}266$ km^2	$x_2y_1-x_1y_3=87{,}665$ km^2
	$x_2-x_1=481.8$ km	$x_3-x_2=-28.5$ km	$x_1-x_3=-453.3$ km
	$(u_2-u_1)_{sfc}=397$ cm s^{-1}	$(u_3-u_2)_{sfc}=-802$ cm s^{-1}	$(u_1-u_3)_{sfc}=-405$ cm s^{-1}
	$(u_2-u_1)_{2000'}=1388$ cm s^{-1}	$(u_3-u_2)_{2000'}=-1882$ cm s^{-1}	$(u_1-u_3)_{2000'}=-494$ cm s^{-1}
	$(v_2-v_1)_{500}=-300$ cm s^{-1}	$(v_3-v_2)_{500}=-566$ cm s^{-1}	$(v_1-v_3)_{500}=866$ cm s^{-1}

TABLE IV-C-7

Centroid values resulting from method 2 computations.

	u_o (cm s^{-1})	v_o (cm s^{-1})	$(\frac{\partial z}{\partial x})_o$	$(\frac{\partial z}{\partial y})_o$	$(\frac{\partial u}{\partial x})_o$ (s^{-1})	$(\frac{\partial v}{\partial y})_o$ (s^{-1})	w (cm s^{-1})
Sfc	-367	-284	-6×10^{-4}	3×10^{-4}	5.6×10^{-6}	-25×10^{-6}	0.14
2000 ft	--	--	--	--	22.8×10^{-6}	-29.5×10^{-6}	0.57
4000 ft	--	--	--	--	16.2×10^{-6}	-28.7×10^{-6}	1.2
500 mb	--	--	--	--	6.2×10^{-6}	-12.6×10^{-6}	3.0

	$(\frac{\partial v}{\partial x})_o$ (s^{-1})	$(\frac{\partial u}{\partial y})_o$ (s^{-1})	ζ_o (s^{-1})
Sfc	--	--	--
2000 ft	--	--	--
4000 ft	--	--	--
500 mb	-0.8×10^{-5}	-2.0×10^{-5}	1.2×10^{-5}

values listed in Tables IV-C-5 and IV-C-6. The results are given in Table IV-C-7. By this method, the value of w at 2000 ft is given by (see IV-11)

$$w_{2000} = 0.14 - 167.3 \times 10^2[(22.8-29.5) \times 10^{-6} + (5.6-25) \times 10^{-6}]$$

$$= 0.57 \text{ cm s}^{-1} .$$

The values of w at 4000 ft and 500 mb obtained using method 2 are given in Table IV-C-7. When the values of w calculated by the two methods are compared, it is seen that minor differences present at the lower levels diminish with height to essentially zero at 500 mb.

b. Calculation of Vorticity. The derivatives needed for the calculation of vorticity at 500 mb also are given in Table IV-C-7. The value of ζ_o computed from these values using (IV-12) is $1.2 \times 10^{-5}\text{s}^{-1}$. With $f_o = 0.85 \times 10^{-4}\ \text{s}^{-1}$, this gives $\zeta_A = 0.97 \times 10^{-4}\ \text{s}^{-1}$.

9. Forecasting of Max Temperature and Convection. Station #2 has been chosen to illustrate the techniques for forecasting the maximum temperature and the possibility of convection because it best meets the conditions required for the techniques to be successful. From the analysis already performed, it was determined at D4-0600 LST that this station is most likely to remain in the mTw air mass through the next 9 h, by which time the maximum temperature should be realized at the station.

It has been estimated that for a station at this Lat in February the insolation for a clear sky can be represented on the Skew T-Log P Diagram by an area of 10° C by 2° θ . However, at D4-0600 LST at this station there is a solid overcast ceiling measured to be 1600 ft AGL. Under such conditions approximately half of the clear sky insolation will not reach the ground. Accordingly, in Fig. IV-C-15, the stipled area represents the net insolation after adjustment for the cloud cover. The mixing ratio line from the surface value of r to the sounding marks the height of the CCL to be at about 940 mb. The intersection of the 940 mb isobar with the pressure-height curve (dash-dot line in Fig. IV-C-16) shows the CCL to be at about 570 m MSL which converts to about 1600 ft AGL (the elevation at Station #2 is 78 m). (A quick approximation can be made by using the Pressure Altitude scale on the side of the Skew T-Log P Diagram). Therefore, the convective activity at this station should simply maintain the cloud deck already present. As a consequence, it is assumed that the

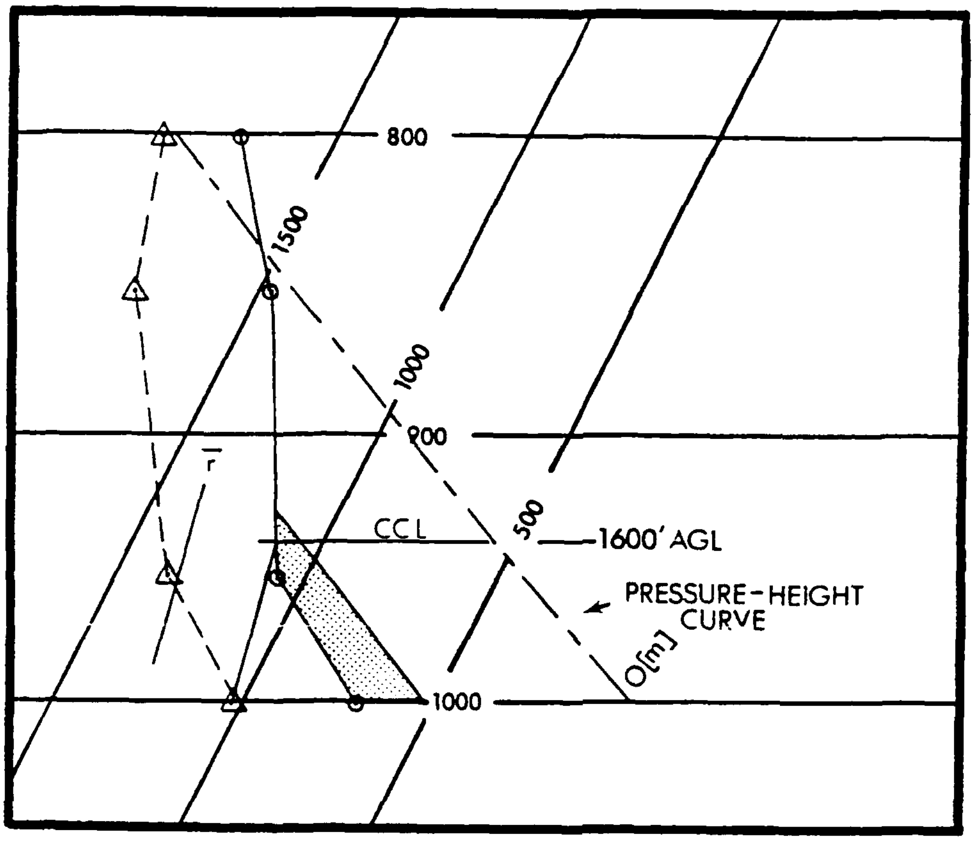


FIG. IV-C-15 Station #2 sounding from 1000 to 800 mb for D4-0600 LST. The dash-dot line is the pressure-to-height conversion curve using the isotherms as a linear scale for height in meters above sea level and the reported mandatory pressure surface heights. The stipled area represents energy added to the boundary layer from solar insolation modified by cloud cover.

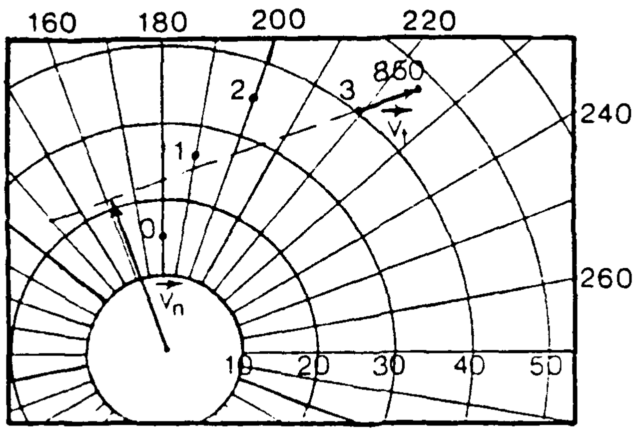


FIG. IV-C-16 A portion of the Station #2 hodograph. The thermal wind vector from 3000 ft to 850 mb is $\vec{V}_t$. The advective wind for the layer is $\vec{V}_n$.

stipled area in Fig. IV-C-15 represents the maximum amount of insolation to be expected. The temperature increase by insolation then amounts to 1.7° C at 1000 mb.

Fig. IV-C-16 shows a portion of the hodograph for Station #2 at D4-0600 LST. Dots represent the winds from the surface to 3000 ft (the gradient level) and the 850 mb wind. The thermal wind, $\vec{V}_t$ from the gradient to 850 mb has a magnitude of 9 kt. The normal wind in the layer, $\vec{V}_n$, also shown in Fig. IV-C-16 has a magnitude of 21 kt. These values along with a forecast period of $\Delta t = 9$ h when substituted into (III-13) give an advective temperature increase, coincidentally, of 1.7° C.

The net temperature increase, 3.4° C or 6° F, from the combined effect of radiation and advection would lead to a maximum temperature of 69° F at about 1500 LST at Station #2. In actuality, the maximum temperature reached was 68° F.

Deep convection should not be expected at this station during the forecast period. The mean mixing ratio line for the lowest 100 mb ($\bar{r}$ in Fig. IV-C-15) would intersect the sounding at a value of θ which, in turn, falls well to the right of the potential temperature line through the maximum temperature for the day. This suppression of deep convection also is indicated by the various stability indices which can be determined for this station at 0600 LST. These are: SI = 1, LI = 2 and TTI = 45. However, light rain showers should be forecast because of the approaching cold front.

10. <u>The Surface Prognostic Chart</u>. Assuming that the surface wave is on the west edge of the map and that it is an open wave, it would be expected to move in the direction and at about half the speed of the 500 mb wind. The direction would be 070° and the speed of 70/2 = 35 kt. Movement for 6 h would give 6 X 35/69 = 3.0° Lat and a 12 h displacement would be 6.0° Lat. The 500 mb wind does become more southerly along the track and the warm front will shift northward so the thermal steering would indicate a shift to the north. The forecast position of the low is placed about 0.5° Lat to the north of the steering wind position. While the movement is good, the location of the low at D4-0600 LST is not known so the forecast low may be off. The warm front can be moved by extrapolation to the new position shown. The forecast cold front is harder to position because continuity is not available. Therefore, use the standard wave and draw

the cold front. The low was estimated at D4-0600 LST to be at 998 mb. The low is crossing about 30 m of contours (Chapter IV-C-8a) at the 500 mb level toward lower values. Thus, a drop of 4 mb at the surface would be expected, maybe to 995 mb. The winds on the west side of the low will be strong and the winds in the warm air will remain strong at the gradient level. The prog map is Fig. IV-C-17.

With the cyclonic curvature the isobars for a 2 mb interval should be about 0.5° Lat for spacing at Station #1 and #2. The wind at Station #2 will be easterly and about 25 kt or greater.

11. <u>The Forecast Made at D4-0730 LST.</u>

a. General. Northern and western parts of area cold, windy, low clouds and poor visibility with snow. Southeastern section warm, with showers but cooling when front passes. (See Fig. IV-C-18.)

b. Station #3. Temperature in the 20's° F. Winds northeasterly at 15-18 kts. Snow and freezing rain reducing visibility to 1/4 mi and making movement of vehicles and aircraft very difficult. The chill temperature will be -10° F.

c. Station #1. Temperature falling rapidly to mid 20's° F with snow and freezing rain. The visibility will be 1/4 mi. Movement of vehicles very difficult. Equivalent chill temperature will be at least -10° F.

d. Station #2. Warm all day, in low 70's° F with southerly winds, low ceiling with showers and thundershowers. Front will pass D4-2100 LST (± 2 hours), then low ceilings and turning much colder, into the 30's° F.

e. Outlook. 24 h of snow, strong winds and temperatures into the low teens. Six to eight inches of snow and deep drifts. Snow may stop by D5-1200 LST, but winds and colder temperatures (10-15° F) will continue over northwestern two-thirds of the area. This is a major cold outbreak.

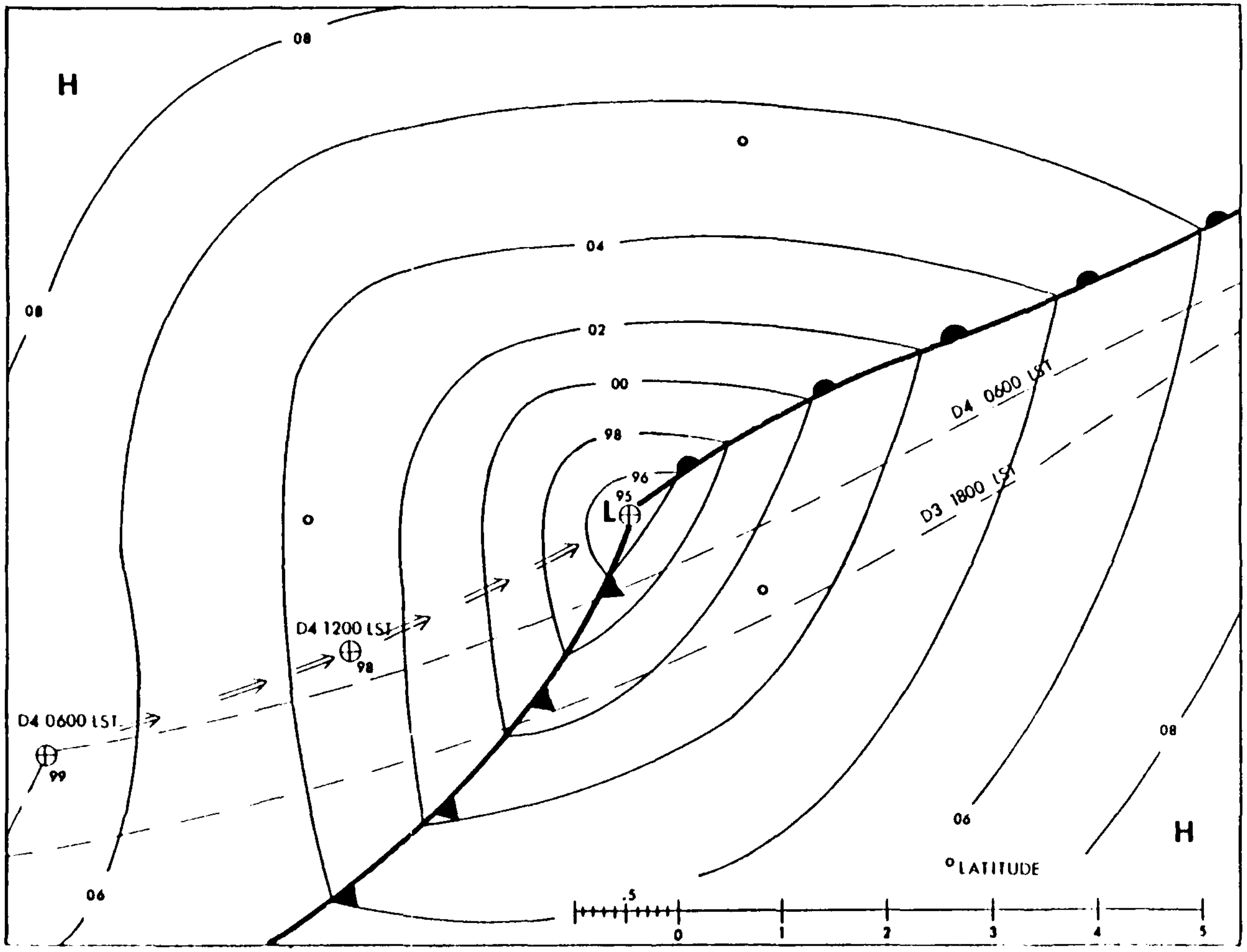


FIG. IV-C-17 Surface prog for valid time D4-1800 LST. Path of low shown by double arrows and past positions of fronts in dashed lines.

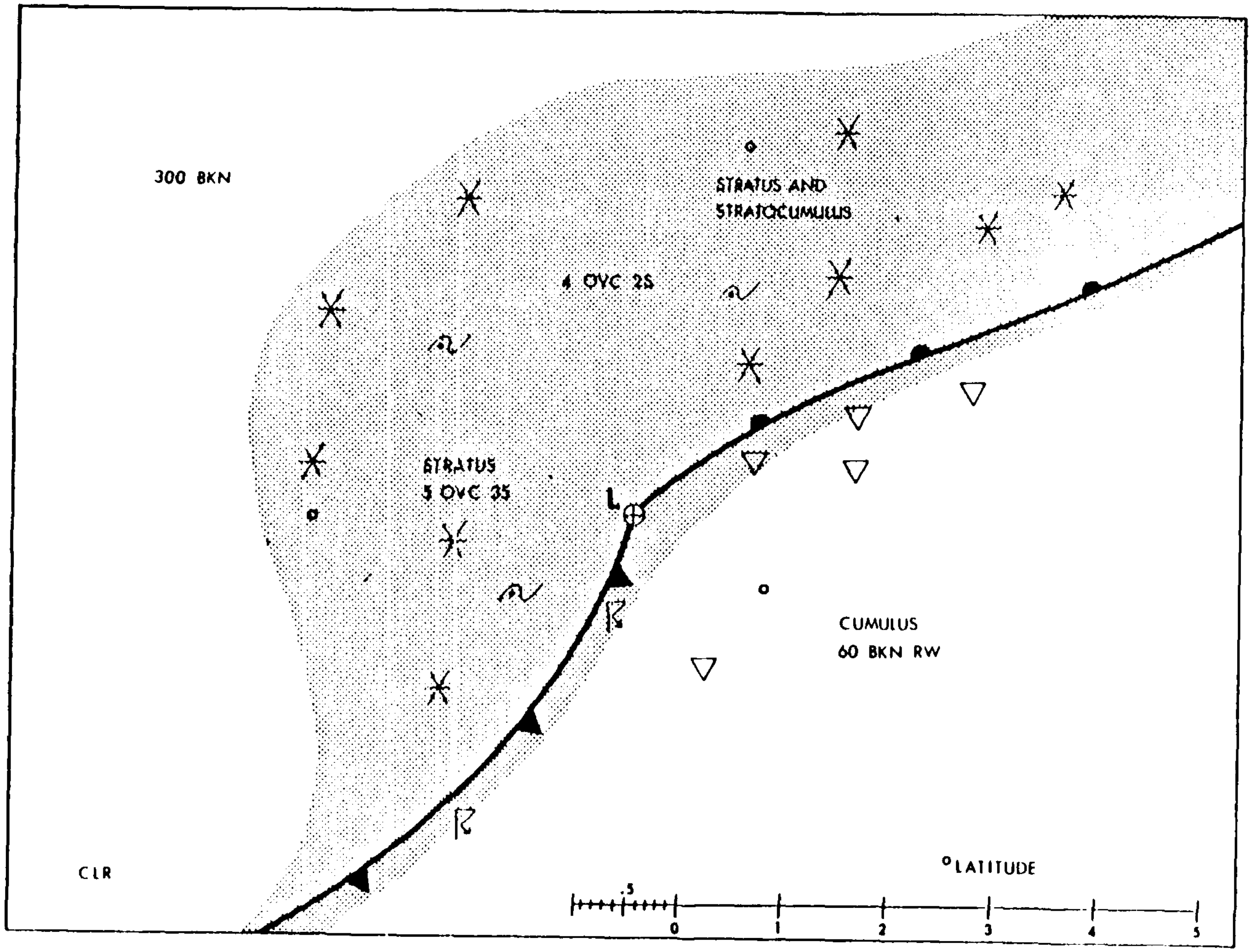


FIG. IV-C-18 Surface prog map for D4-1800 LST with forecast for clouds and weather under the shaded area.

CHAPTER V

SINGLE STATION ANALYSIS IN THE ARCTIC AND TROPICS

PART A GENERAL CONDITION

1. Arctic. The arctic will be considered the area within the Arctic Circle, plus areas of permafrost and areas included in the "E" type of Köppen climatic classification. The arctic (and antarctic) areas have extreme sunshine patterns. In the winter the sun rises in the southeast and sets in the southwest when it makes it above the horizon. In summer, if the sun sets, it will be in the northwest and will rise in the northeast. Because of the sunshine patterns, winter and summer have marked differences. Phenomena occurring in the arctic that are not common in the mid-latitudes are whiteout, looming, and ice fog. Any forecaster going into the area should study the meteorological conditions, the terrain, and geography of the area before arrival. Cold weather survivial training is desireable in any condition, but a must if the team is to move into a hostile area.

    a. Air Masses in the Arctic. The northern hemisphere arctic air masses are cA, mA, cP, mP and on occasions mT.

        1.) mT air mass. Under certain conditions the mT air mass will invade the arctic. When it does in winter the weather usually is very disagreeable. The warm, moist air causes heavy precipitation and thaws some of the ice on the roads so they become slick (cold ice is not slick and makes an excellent road surface). A slow moving 500 mb trough with large amplitude can bring the mT air into the arctic. In summer, additional rainfall is about the only problem with mT air.

        2.) mP air mass. The mP air mass is a frequent visitor aloft in the winter, but in summer it is the prevailing air mass at the surface. The warm, moist mP air rides over the colder cA air in the winter, causing clouds and snow. The leading edge of the advancing mP air is called a "Trowell" by the Canadians.

        3.) mA air mass. The mA air mass is a transition stage as the cA air moves from land to open water. The cA is soon modified

to a mP air mass.

4.) cP and cA. Some meteorologists do not separate these two air masses. They have the same characteristics, except some of the air is colder which, at least for convenience, may be called cA. Both air masses are shallow, cold, and have low mixing ratio values. The cA air mass does not exist in the summer, except on the snow-topped mountains.

b. Upper Air. The arctic area is to the north of the polar jet streams (especially in winter) except when the top of a ridge may extend well to the north. Therefore the winds at 500 mb often are weak, but they may be very strong on occasion on the top of the ridge. The Stratospheric Polar Night Jet does develop in the winter at 50 km above the surface. This jet does not seem to have any day-by-day effect on the surface weather but would be important for aircraft operating at the altitude.

c. Fronts. The boundary between air masses is known as a front. The front with the cA air mass on its cold side is called the arctic front. Frontal waves will form on the arctic front as on other fronts. Because of the weak winds aloft and the shallow air mass, the wave on the front does not move rapidly. Because the air mass is shallow it does not cross over mountainous areas, but tends to go through mountain passes or go around the obstruction. The location of the front is usually marked by a band of clouds and snow. Since it moves slowly some time may occur before it passes the next station. For conventional analysis the forecaster must keep track of the time the front passes each station because often after it passes the winds and temperature may be more affected by terrain than the conditions of the new air mass, thus the front is not identified easily.

d. Radiation. Because of the long nights during the winter, which usually have clear skies, the long wave radiation is a factor. On many soundings the inversion for the arctic front can be identified, and an additional very cold layer at the surface is the radiation inversion. The radiation inversion may be destroyed very rapidly when a cloud layer moves over the station. As a general guide a cirrus overcast

will cause a 10° F rise, a middle cloud overcast will cause a 20° F rise and stratus will cause the temperature to rise 40° F in less than 2 h. The arctic air mass will still be there but the extreme cold produced by the strong radiation losses will be modified.

Because the very cold air drains into a valley, the hillsides are always warmer than the valley. Also, the hillsides are windier than the valley where the dense cold air is calm.

e. Ice Fog. At temperatures colder than -22° F ice fog may occur. There are two types of ice fog, pure ice fog and induced ice fog. Induced ice fog occurs when additional moisture and/or pollutants are added to the air, such as from a smoke stack, auto or aircraft exhaust, or even a moose walking across the tundra. This induced ice fog occurs around inhabited areas and may be initiated at -22° F but almost always at -40° F. The fog is shallow and the top of hangers may be above it. A pilot is often misled because when he flies over the runway he can see everything. (looking downward through only a few feet of ice fog) but as he lands, his horizontal visibility suddenly may become nil.

The ice fog is most dense during the day when more activity is occurring (cars moving, aircraft operating, etc.). If activity decreases the ice crystals will settle and at about 0600 LST the visibility is good and will remain good until the first aircraft operation leaves a contrail on the runway that soon spreads across the area.

True ice fog does not form until the temperatures drop to about -50° F. Of course, around cities, the induced ice fog will already have formed as the temperature decreases. Regardless of the temperature the first day after an arctic frontal passage, ice fog does not form.

f. Blowing Snow. Because blowing snow reduces the visibility and obscures land marks from the air, it is a condition that must be considered because it is so dangerous to anyone outside. At some camps ropes are stretched between buildings so that a person may hold to them and not lose their way in the poor visibility or be blown away by

the strong wind. The forecast procedure is the same as anywhere else; look for loose snow and stong winds. Blowing snow is not a whiteout.

g. Thunderstorms. Thunderstorms occur over the inland areas in the summer season. The surface temperatures may reach 100° F and the lower level air is unstable. Terrain is available for orographic uplift. In inland Alaska, a SI of less than or equal to 4 is critical for thunderstorms to form.

h. Surface Winds. During the winter months the surface winds in the interior of land masses are weak, but along the coast, the winds are stronger (150 kt or greater may occur in major storms). In the summer the interior winds are usually less than 15 kt and the coastal winds remain strong but not as strong as winter.

i. Additional Information. A number of studies have been made on arctic climate and weather. Some, but by no means all, are listed: Director of Met. (1960), Rubin (1966), Petterssen et al. (1956), Gavrilova (1966), Reed et al. (1956 through 1959) and Weller and Bowling (1973).

2. Tropics. The tropical area is usually defined as that area between the Tropic of Cancer and Tropic of Capricorn. However, it may be extended in a meteorological sense to include all of the "A" area of Köppen's classification and all of the area which usually is covered with mT or cT air.

a. Air Masses in the Tropics. The air masses are mT, cT, mP and cP. In the literature some authors mention some others. One article describes cA air as far south as 15° N. E has been used for Equatorial and M for monsoon air masses.

1.) mT air mass. Since most of the area on both sides of the equator is covered by ocean, it is logical that the most extensive air mass in the tropics is mT. The warm, moist air usually has clouds almost every day and often has showers and thunderstorms.

2.) cT air mass. Since there are not large land masses in the true tropics (Africa being an exception), not much cT air exists there. There are arid areas which may be classified as a "B" climate by Köppen, but the surface air is usually mT. The cT

air is on the fringe of the tropics in the desert areas about 30° N and 30° S. Aloft, above the mT air, the airmass is dry and potentially warm and when brought to the surface has the characteristics of the cT air mass.

3.) mP and cP air masses. The polar air from mid-latitudes in the winter season will penetrate into the tropical areas. The temperature of the polar air is warmed rapidly but the native of the tropics notes the colder air.

b. Severe Storms. Some of the thunderstorms do become strong and do have strong downrush winds. They are usually in clusters or lines which can be identified as compared with the isolated thunderstorm.

c. Fronts. Cold fronts from both the Northern and Southern Hemispheres will invade the tropical areas and occasionally may cross the equator. The surface temperature change will be small but the dew point changes, wind shifts, and rain showers are there. The temperature changes at 850 mb are usually larger than temperature changes at the surface. The discontinuities may be called fronts, shear lines, convergence zones, or squall lines, but they must be analyzed because of the weather they cause.

d. Pressure Changes. For many years forecasters have attempted to use the pressure changes to predict weather in the tropics. In general, a 24 h pressure change was used so as to eliminate the diurnal effects, and also used was a 3 h period vs a 3 h period of the same time the previous day. In general (excepting tropical storms), these have not been too successful (Landgren and Leies, 1945). Rises and falls of several millibars may occur over a large area for reasons not known.

e. General. Of all the elements of the tropics the rainfall is the most important to the welfare of the economy of the country. Note that the temperature stays within well-defined ranges, the winds are not too strong, hail occurs in higher plateaus and tropical storms develop mostly over the oceanic areas. But to an agrarian economy the rain is the most important of the weather elements. Too much rain causes floods and too little means no crop. For military oper-

ations, the clouds, rainshowers and thunderstorms and severe storms, i.e. hurricanes, will be the main forecast problems.

Atkinson (1971) is available as an excellent starting point for study of general conditions in the tropics. Additional references are Anthes (1982), Palmer et al. (1955, Vol I; 1961, Vol II), Riehl (1979), Shaw (1978), Simpson and Riehl (1981), Sutcliffe (1958), Weikes (1957), and Winn-Nielsen (1979). Much material concerning convective activity in the tropics has come from "GATE" and can be found in the recent journals.

PART B PRACTICAL ASPECTS

1. Arctic. Single station analysis is always helpful in the arctic (and antarctic). Because of the great expanse of uninhabitable land and water, large areas have no reporting stations, thus SSA is a continuing requirement. SSA techniques should be used at all stations, especially the stations on the edge of the reporting network. The terrain in much of the Arctic land area is mountainous so the cold air is impeded in its movements. Each major valley has its own synoptic problem and may have only one or, at most, three reporting stations. Analyze each station in each valley by SSA then join the several analyses to make one good area analysis. The concepts of SSA do not need to be changed for the arctic; however, there are some complicating factors which the forecaster must consider. The most important factors are discussed below.

   a. Bodies of Water. In the winter, lakes and the Arctic Ocean are covered with ice which will be several feet thick. There is more heat transfer to the overlying air mass from the Arctic Ocean than from the land masses, thus the coldest air is over land (Siberia). From the standpoint of a day to day forecast the frozen lakes and ocean may be treated as a land surface without any transfer of water vapor or heat from the surface to the air. There are a few hot springs that will cause interesting micro-climates.

   In winter, along some coastlines the water does not freeze. A rapid exchange of heat and moisture occur when the cold air flows from land to water. With light winds "sea smoke" will form over the water. An onshore flow will cause cooling of the warmer and

moist air from the ocean causing advection fog which may be both deep and thick.

During the summer the ice on the lakes and rivers melts, but the water stays colder than the land. The forecaster must know his local area especially during the summer.

The SSA data will be influenced by the land-water distribution so the forecaster must consider these modifications. The effects cause the same problem as in mid-latitudes. The data may need to be adjusted to eliminate local effects to establish the more general synoptic pattern.

b. GLW. Selecting a representative GLW requires some judgement, especially in the winter seasons when the radiation inversion and cold air drainage are both very strong. The surface will be calm and very light winds exist up through the cold layer of surface air. A wind in a layer above the radiation inversion but below the arctic front would be ideal but often such a level cannot be determined because the radiation inversion and the frontal inversion blend together or the transition layer is so thin that no wind is recorded in the layer. There is no rule; pick the wind that seems to be the best for the day.

c. Shear Vectors. Since the arctic front is shallow the shear from the GLW to 700 mb wind is used to orientate the front. However, the terrain may have more influence on the location of the front than the meteorological dynamics. The shear vectors at the 700, 500 and 300 mb levels are good for orientating isotherms at these levels. Check the advection because warming or cooling aloft will be reflected on the surface the next day.

d. Tropopause. The tropopause will be lower in the winter than at mid-latitudes. There are reports from some arctic stations which indicate that on occasion the tropopause may be on or very close to the surface. It may cause some problems when drawing the temperature patterns at the 500 mb level, and often will be a problem at 300 mb.

e. Vertical Motions. During the arctic winter the cold surface air and lack of sunshine prevent convective activity. Vertical motion

is confined to terrain effects.

1.) cold air drainage. The air on the hillside cools by outgoing longwave radiation. The cold air drains into the valleys, following the same path that water runs down the hillside. This cold air collects in the valleys. A person can note the temperature change when climbing the hill or descending into the valley. The drainage winds are weak, maybe 1 or 2 kt. The drainage will continue throughout the 24 h day. The upslope wind of mountain valley systems in the other parts of the world does not occur because of the lack of sunshine during the day. The single station forecast for these drainage winds is not a problem because they are weak and are always there in the cold dark months.

A special case of local importance is the glacial wind. The air over a glacier becomes even colder than the drainage air discussed in the previous paragraph. This air when restricted to a narrow valley with a steep slope <u>may</u> develop winds even exceeding 100 kt. At the entrance of the valley the glacial winds flow out (direction of the orientation of the valley) and once out of the valley spread into the flat area and lose speed. Only the local area in the valley and just outside the entrance of the valley are affected. The three needed imputs are the temperature difference between the ambient air and the air at the glacier, and the slope, shape, and resistance to air flow of the canyon. In additon the wind direction aloft, at glacier height, must be considered because it may hinder the cold air drainage or encourage it. A local forecast study would be needed to relate these factors to provide a forecast.

2.) upslope winds. A wind flow encountering a mountain range may be lifted at all levels. The mountain wave is a common occurrence in the arctic area because of the stable air and the available mountains. The downslope winds on the lee side of the mountains may be very strong on the mountainside. However the foehn winds usually will not penetrate the very cold air on the floor of the valley. The forecast using SSA is easy when the station

is on the windward side because the wind profile and atmospheric stability is known. If on the lee side, the forecaster will need to extend his 700, 500, and 300 mb analysis upstream past the mountain range. The visibility is good on the lee side and the rotor clouds and lenticular clouds may be seen so the forecaster knows the wave is occurring. Should the downslope winds be strong enough to penetrate the cold air mass, warm spots, known as foehn islands will be observed.

3.) lifting and divergence aloft. During the winter the tropopause will limit any forced lifting over the terrain of any layers. The winter tropopause is low and will cap any vertical motion. During the summer the tropopause is about the 300 mb level. Air will become unstable from surface heating or uplift over the terrain. The unstable air will rise to the tropopause which is the level of divergence.

f. Polar Front Occulsions in the Arctic. The mP fronts have waves that move across the Atlantic and Pacific Oceans. The low will cause the mP front to form occlusions. The surface low at the end of the occulusion may shift north of the jet stream and a new low form at the point of occlusion which is south of the jet stream and toward the eastern side of the trough of the major wave. When the new cyclone forms, the occlusion may break loose and be carried northward in the southerly flow on the east side of the major trough. With a trough of large amplitude the occlusion will be carried into the arctic area. When it encounters the cold cA or cP air in the Arctic the occlusion moves aloft above the arctic front. Thus another occlusion has formed. From the forecaster's viewpoint it is better to analyze the old occlusion as a front aloft, above the cA air and not try to burden the analysis with another occlusion. The thermal structure of the front aloft (the old occlusion) is weak because it has mP air on both of the occlusion. Because of the small ΔT across the front it will have a steep slope. At greater elevations, (500 mb), the mT air from the warm sector of the original occlusion

may be present. If so, the ΔT across the front is greater and moisture from the mT air will cause more clouds and snow.

From a single station analysis standpoint the following conditions occur when the old occlusion is in the vicinity of the station.

1.) no change of surface air mass.

2.) because of the cloud cover the surface temperature may change (see previous radiation inversion discussion).

3.) clouds may be on both sides of the occlusion aloft. The approach of the clouds follows the same pattern as the clouds preceding a warm front. Rapid clearing indicates the front is past. However the front may pass and the clouds may remain for a day.

4,) precipitation may fall, usually as snow.

5.) the surface pressure will fall as the occlusion approaches and rise after passage.

6.) the heights at 500 and 700 mb will lower, then rise after passage. To observe the height change compute the 12 h ΔH and 24 h ΔH.

7.) the winds aloft will make a cyclonic shift when the occlusion passes. The occlusion is in a minor trough moving northward over the major ridge. The wind shift may be from the southeast to the southwest and a shift to north wind may never occur.

8,) the temperature at 500 and 600 mb will increase then decrease. A 24 h ΔT may be from 2°C to 10°C. The occlusion with the 10°C ΔT will still have several days of existence.

9.) the speed of the occlusion is the speed of the 500 mb level wind.

10.) the shear vectors at 700 and 500 mb will indicate the location of the warm and cold areas which will be warm ahead of the minor trough and colder behind.

11.) the forecaster must be aware that the occlusions may come past and be alert to identify them because they cause most of the fall, winter and spring precipitation and cloudiness.

g. Ice Jams and Floods. The ice on the rivers will be at least 6 ft thick in the springtime when the thaw starts. In addition to the

top ice, the river has frozen from the bottom upward because of the permafrost. When snow melt or rain water runs into the river and raises the water level, the top ice will separate from the anchor ice and will start to float downstream. The ice pans, 6 ft thick, 100-200 ft long and 30-40 ft wide cannot turn the bends in the river, thus they jam together, holding back the upstream ice. Under pressure the ice pans will fuse into one solid mass. The ice-jam will soon restrict the flow of water thus causing the water to flood over the area above the jam. At this time it is usually suggested that the ice-jam be bombed. This is a mistake because the anchor ice on the bottom is blasted loose so then twice as much ice is available to make a jam. A forecaster can determine the years that a jam will form. When the headwater area warms and melts snow before the lower reaches of the river thaw, ice jams will form. If the lower reaches thaw first then the ice is gone and there may be high water but no ice jam floods will occur (Henry 1965).

2. Tropics. As in all meteorological forecasting the local conditions have an influence. The forecaster must know the terrain and geography of the area. The windward and lee sides of the mountains, and even smaller hills cause uplift or downslope winds which may cause clouds, showers and foehn conditions. The vegetation of the land mass changes the albedo which leads to different heat effects from the sunshine. Lakes, curvature of coastlines, rivers, swamps and all other features, even man-made cities have an effect.

The climatology of the tropics indicates smaller ranges of temperature than in other areas. The temperature usually stays close to the monthly mean. The surface winds have more persistence than in other areas of the world but climatology alone cannot be used for a good forecast. It can be used to prevent an unlikely one. Rainfall is the large variable, and monthly means are a poor guide (See Appendix, Part O). Climatology is very useful for determining the rainy seasons of the area. A knowledge of the climatology is very important for meteorological analysis and forecasting in the tropics, as in all other areas of the world.

A forecaster cannot be successful without a knowledge of the terrain, ecology, and the geography of the area.

The tropical areas are made for SSA. Even in the best of times the stations are isolated. The x-t cross-section was developed in the tropics. A close watch must be maintained to notice small changes. The tropical area has been described as a diurnally-persistent region. This is an erroneous concept. A persistence forecast will not suffice.

a. Permissive vs Suppressive Days. Almost all the rainfall in the tropics comes from mesoscale systems (showers and thunderstorms). A permissive synoptic system allows or even encourages the mesocale systems to operate, thus much precipitation. On other days the synoptic conditions are suppressive so the mesoscale systems do not develop, thus little or no rain. The atmosphere is in a permissive condition for 3 or 4 d in succession with a shorter suppressive period. During the dry season the synoptic scale conditions suppress the mesoscale systems. However, some strong mesoscale systems do break through and heavy rains occur. The dry season thunderstorm gives more rain than the wet season thunderstorm. The wet season just has many more thunderstorms. Some authors refer to the permissive days as "+" days or disturbed, and the suppressive days as "-" or undisturbed days. More on forecasting by SSA is below in c.

b. Shower Size. The rain showers are arranged in patterns so that only about 12-15% of the area will have precipitation any one day. The cells are about 35 mi apart and are about 15 mi in diameter. A forecaster should know that most of the area does not have thunderstorms even if a thunderstorm occurs at his station. For more information about the nature and variations of rainstorms, read Henry (1974), "The Tropical Rainstorm" and Part O of the Appendix.

c. Conditions to Watch. A quick look at a time series of tropical data indicates a sameness which is deceiving. Because of this many forecasters do not appreciate the value of the rawinsonde data in the tropics. A small change may be enough to cause a major weather event.

   1.) rawinsonde data. These data are useful for determining the

trade wind inversion, precipitable water, θ_E, and stability. Stability will be discussed below. The trade wind inversion separates the moist mT air at the surface and lower levels from the drier air aloft. The easterly winds in the lower levels may change to westerly above the inversion. The inversion suppresses cumulus development. Furthermore, the dry air aloft mixes with any cumulus clouds that penetrate the inversion and may evaporate them. The precipitable water is a good indicator of the moisture aloft. In many locations it can be related to the probability of precipitation. When the values of precipitable water are low (< 1 in) precipitation is unlikely; values larger than 2 in in the tropics does not mean rain will occur, but the chances have increased.

The lapse rate of equivalent potential temperature will show differences from day to day better than the temperature lapse rate. A change in $\partial\theta_E/\partial z$ relates very closely to a change of stability and the changes from sounding to sounding should be noted. These changes are the indicators of "+" and "-" days. Atkinson (1971) describes the variations of θ_E with height very well with several diagrams.

In the presence of the trade wind inversion and with very few means to provide lifting, organized release of latent instability is impossible in the tropics. Large values of θ_E in the boundary layer coupled with horizontal convergence is necessary to initiate convective overturning.

2.) stability and shear. Krishnamurti (1985) states that in a barotropic atmosphere, an increase in time of horizontal cyclonic shear at the 850 mb level tends to decrease stability in time. From a SSA viewpoint the horizontal cyclonic shear is hard to determine anywhere, but especially in the tropics. A time series of wind observations can supply data for an estimate of horizontal shear when several assumptions are used. The main assumption is that the previous wind has translated some direction and distance and still maintains its direction and

speed. In some cases this assumption may be met and a horizontal wind shear can be computed. In the mid-latitudes the pressure system is translated. The continuity of a pressure system is better than the continuity of a tropical wind field.

The concept of a barotropic tropical atmosphere like many of the earlier beliefs about the meteorological conditions in the tropics, is being phased-out because in fact, the tropical atmosphere is baroclinic. The baroclinicity is not as strong as in the mid-latitudes but there is enough to cause sloping systems and shear vectors (thermal winds).

The gradient wind, geostrophic wind, and thermal wind all have the Coriolis parameter ($f=2\Omega\sin\phi$) as a factor. ϕ is the latitude and at the equator, $\sin\phi=0$, which means that the equation cannot be applied at the equator. However, the wind does blow at the equator. Between 5° N and 5° S the geostrophic relationship does not hold too well, but the Coriolis effect is active. Any flow of air coming across the equator is turned 90° by the time it has reached 5° N or 5° S.

At latitudes outside of $\pm$ 5° Lat the concepts of thermal wind, advection, and slope of systems may be used, and with more confidence farther from the equator.

Krishnamurti (1985) comments on the baroclinity of the tropical atmosphere and the shear in the vertical. An easterly (shear vector pointing eastward) vertical shear of 30 m s^{-1} from 850 to 500 mb will imply dense convective clouds in the shear area.

The shears from 850 mb to 500 mb along with the TTI is one of the better methods of forecasting the location of rain-showers. The temperature gradients are weak but, when the atmosphere is (in the rainy season) conditionally unstable, does <u>not</u> have the tropical inversion, and moisture extends to high altitudes, not much differential advection is needed to let the cumulus clouds grow. With one RAOB station the direction of the most activity can be determined and, if three or four

RAOB stations are available, the area which will have the rain can be identified from the morning soundings. The shear of shears indicates the permissive and suppressive areas. In Venzuela this method has been so successful that the stability chart with the shear of shears has become the main chart for shower and thunderstorm forecasting.

Atkinson (1971) has pointed out that stability indices have limited use in the tropics. Of course no thunderstorm was ever caused by an index in the tropics or elsewhere. The usefulness of indices are to designate areas that need to be considered first because they are more unstable at the time the observation was taken. The differential advection indicated by the shear of shears will change the stability. Diurnal heating is another factor which changes the stability over a land mass.

3.) The sea breeze. In many parts of the tropics the sea breeze has dictated a way of life. The afternoon siesta is during the time period between noon and the arrival of the cooling sea breeze. The sea breeze is a flow of air from the sea (ocean, lake) whenever the adjoining land surface become warmer than the water. A difference of 10-15° F will usually start a sea breeze and the greater the temperature difference the stronger the sea breeze and the farther inland it will penetrate. The GLW may be in a direction to support or stop the sea breeze. The cooler, but moist sea breeze is heated as it flows over the hot land and may become unstable so that showers develop. Other factors such as terrain, plant coverage, soil color and moisture, existing stability or inversions all tend to affect the sea breeze and the following events. Atkinson (1971) devotes six pages to the structure of the sea breeze.

The forecaster with only single-station data would note that the sea breeze arrived by the surge of wind speed and maybe a change of direction. The sea breeze direction usually is normal to the coast at the coastline but turns toward the right as it progresses inland (NH). Bluffs, hills, coastal islands

deflect the sea breeze and change both the speed and direction. An increase of the relative humidity and T_d always occurs and some cooling may be observed. The cooling may be indicated only as a decrease of the heating rate, depending upon the time of day.

From the operational SSA forecaster's viewpoint, the possibility of clouds, showers, and thunderstorms becomes the main problem. The air of the sea breeze is heated from below by transfer of heat from the hot surface. Since some time is required for the air to be heated to instability the beach does not have many showers, but the showers occur inland. If the sea breeze is moving upslope, it will help reduce the stability. In most cases the sea breeze is an upslope wind.

The rawinsonde data prior to the arrival of the sea breeze needs to be considered. The direction of the GLW will indicate the resistance to the sea breeze and the convergence that could occur. Any existing inversions could suppress the lift and keep the convective activity from occurring, so the sounding must be studied. The increase of the T_d may lower the CCL to a level below the inversion. Small cumulus clouds may be the only resulting activity.

Once thunderstorms start, their cold outflow will meet the incoming sea breeze and the convergence will continue to generate more thunderstorms. This is especially true when a coastal mountain range is present because the cold outflow will move in one direction, downhill, increasing the intensity of the convergence and thunderstorm quasi-front.

The land breeze which blows at night from the cooler land to the warmer water usually is not a problem. However, in some places it does cause a few thunderstorms over the water during the pre-dawn hours.

4.) the monsoon. The monsoon is a seasonal change of the wind direction. The wind shift usually brings a rainy or dry period depending upon the wind direction. The cause of the monsoon is attrib-

uted to the large temperature change on the continents as compared to the oceans. Many parts of the world have some monsoonal characteristics but the best developed monsoon is around India and southeast Asia.

The climatological date for the arrival of the monsoon is known for most places in the world, but the actual arrival may be several weeks from the mean date. The forecast of the "burst of the monsoon" is difficult even when all data are available. However, Krishnamurti (1985) shows a case for which a computer model made a good forecast for both the start and a break in the monsoon.

The forecaster with only SSA should know the climatic date, so the event would not be unexpected. Dr. Phanindramohan Das* (personal communication) states from his experience of forecasting in India that the arrival of the monsoon at a coastal station was watched, so that the movement inland could be forecast. From a SSA standpoint, often heavy thunderstorms would occur two or three days before monsoon rains arrived. The movement of the monsoon rains across an area is not continous, but a series of surges, each extending farther than the previous one. He was of the opinion that the 850 to 200 mb shear vector became stronger just prior to the arrival of the monsoon.

Krishnamurti (1985) comments on the large scale atmospheric structure and identifies three indicators of the onset of the monsoon: 1, a rapid increase of daily precipitation rate, 2, an increase in the vertical integrated humidity (over a large area), and 3, an increase in the kinetic energy, especially in low level flows. Of these the SSA forecaster could note indications of all three, depending upon his location. Number 1 would identify easily, number 2 would show on the radiosonde sounding by persistence, and number 3 would show only if the SSA data were in the correct relation to the wind pattern.

---

*Dr. Das is a member of the faculty of the Meteorology Department at Texas A&M University.

5.) ITCZ. The Intertropical Convergence Zone is the source of much of the thunderstorm activity in the tropics. The ITCZ usually is a broad area of showers and the zone is too wide to mark on the map with only one line. It does have seasonal travels, following the sun with a lag time of about one month. Atkinson (1971) divides the ITCZ into two types, the monsoon trough and the trade wind trough. When it is well north of the equator the trade winds from the southern hemisphere cross the equator and are turned to become westerlies. This is one type of monsoon, as the winds of the area change from easterly trades to westerlies in what Atkinson (1971) calls a "buffer zone".

In the area of the ITCZ (by any name) the temperature inversion and dry air aloft of the trade wind area of the tropics are missing, or so weak that they are hard to identify. This area of the tropics does approach the barotropic condition.

The SSA will show the wind distribution and the moisture aloft so the forecaster can identify the system. The forecast problem is the same with one or 10 radiosonde stations in the area. With the 10 stations the trough may be located better and its width identified better, but the SSA will indicate the important features, the moisture aloft, and the θ_E lapse rate.

6.) cold air aloft. In both the middle and upper troposphere cold pools of air with a cyclonic circulation do drift through the tropical areas. It has been known for years that the cyclones cause rainfall and are related to the tropical storms. Climatology and knowledge of the existing models has been stressed throughout this report. The analyst and forecaster must be able to position his station in the appropriate model. Atkinson (1971) describes tropical synoptic models in Chapter 7. The effects of terrain and geography will be superimposed upon the model.

From a single station viewpoint these small (2000 km diameter) cyclones in the upper troposphere can be identified only when their circulation and temperature patterns influence the data.

They are difficult to locate even with all rawinsondes in the area available. In addition to the rawinsonde data the clouds and weather must be observed and used. Since the same type of clouds and weather occur with other conditions, they are not sufficient by themselves, but may be used to support the rawinsonde analysis.

Atkinson (1971) shows a proposed mean trough and ridge model which he presents as Figs. 7-25 and 7-26. The troughs extending NE-SW over both the Pacific and Atlantic Oceans in the Northern Hemisphere have been named Tropical Upper Tropospheric Trough (TUTT) by Sadler (1976a, 1976b, and 1978). The August position is the northern extent of the TUTT. A similiar pattern occurs in the Southern Hemisphere in the summer season. For a more complete description see Atkinson (1971) or Sadler (1976a, 1976b, 1978). The trough is not stationary or continous as shown in the mean, but shifts from day to day and even breaks. Sections of the TUTT may take a N-S orientation as shown by Sadler (1976a).

a.) With a series of rawinsonde data the forecaster can locate his position in relation to the troughs and ridges by the direction of the winds between the 300 mb level and the tropopause. The TUTT is a cold trough and the shear vector between the 200 and 300 mb winds will indicate the direction of the colder air. The NE end of the TUTT has a greater temperature contrast than the SW end so the shear vector will be stronger in the NE.

b.) The pools of cold air enter at the NE section of the TUTT and in general drift toward the SW in the summer. These cold pools usually have a closed cyclonic circulation, and of course, are in the TUTT. Their approach may be identified by 12 and 24 h temperature changes, 12 and 24 h height changes at 300 and 200 mb, and the winds at the upper levels becoming stronger and changing direction as the low approaches and passes. The shear vector is useful

to locate the direction of the cold air, which is in the cyclone.

c.) The cloud pattern of a cold cyclone in the TUTT and its identification from a series of surface observations is not possible because the visible celestrial dome is too small. However, when a cold cyclone is suspected by wind changes the cloud pattern may help to confirm its presence. Often during the first three or four days of the cyclone while in the mid-latitudes, no clouds, or only a few clouds will exist. A band of cirrus in the east winds north and northwest of the cyclone is one indicator. As the cyclone moves southwest and moves over the surface trade winds rainshowers and the CB clouds form to the southeast of the low aloft. However rain showers occur at other times with different synoptic patterns.

d.) In the fully-developed case, tropical storms develop to the southeast of the low aloft. In other cases, especially to the far western end of the TUTT, extensive rain may occur. The intensity of the surface weather caused by the cyclones is difficult to forecast with all information available and if more than 1000 km from a single station a tropical storm will not be identified in its initial stages.

e.) It should be noted that the TUTT concept is not considered by all authors (Riehl, 1979) when describing the cold air aloft. SSA will identify the cold air cyclones aloft by any name one wishes when they come over the station.

f.) The origin and behavior of the mid-tropospheric cyclone is well-described by Atkinson (1971). The SSA will locate this cyclone's approach by ($\Delta H/\Delta t$), ($\Delta Th/\Delta t$) and by ($\Delta T/\Delta t$). Both 12 and 24 h differences should be used. The shear vectors at 500 and 400 mb (see Table-B-1) will indicate the direction of the cold air at those levels. The cyclone will be located to the left of the wind direc-

tion (NH). Stronger winds and changes of wind direction around the cyclone can be seen on the time x-section. The mid-troposphere cyclone travels an erratic path and may persist for several weeks, causes much rain and sometimes development of surface cyclones.

g.) During the winter season cold pools of air become separated from the westerlies and drift into the tropics. They may cause much rain. The SSA will identify the cold cyclones when they are affecting the area.

8.) hurricanes and typhoons. A forecaster with only SSA should be able to identify an approaching storm 24 to 48 h before arrival. Cirrus and cirrostratus clouds will often form in a banded structure. The bands have the appearance of converging toward the horizon, in the direction of the storm. Soon afterward, the cirrus will increase and the bands no longer can be identified. Along the coast the increased surf at fairly regular intervals is a precursor. The crest line of the wave is approximately normal to the azimuth of the storm direction. The waves move faster than the storm so the waves from the storm arrive before the increase of the wind. As the storm approaches the pressure will fall and the winds will increase. The low clouds and rain will develop. The GLW will become strong and the storm should be in the direction normal to and to the left of the GLW.

Using SSA from only one station the direction of the storm can be determined but not the distance. With two or three stations, the center may be located by triangulation. However, with one station the track may be established in relation to the station. The storm will be approximately normal to the GLW and the 850 mb wind. The change of direction of these winds will indicate lateral movement of the storm. If the wind direction does not change but the speed increases the storm is approaching or becoming larger or both. The pressure change at the station will help determine if the storm is approaching the station. The barometric pressure change may indicate a diurnal

pressure change or a change of intensity of the storm or both. The pressure and wind speed <u>cannot</u> be used as an absolute measure to determine the distance to the center, because the strongest wind and the lowest pressure of the storm are not known. If the eye should pass over the station then, for the departing storm better estimates of the distance to the center of the storm may be made using the climatological profiles of pressure and wind of tropical storms.

Again, the climatology of tropical storms should be used because the winds are not uniform in the storm (usually stronger in the right, front quadrant). Also, the winds are stronger in the rain bands. A few tornadoes may be associated with the storm, but if the storm should curve toward the south the number of tornadoes usually increases. Along the coast the forecaster needs to know the terrain because a related hazard is the storm surge. The depth of the storm surge is a function of the slope of the beach, the contour of the coastline, and the speed of the wind.

CHAPTER VI

STATISTICAL SINGLE STATION FORECASTING

PART A CONCEPTS

The direct utilization of information from the surface and RAWINSONDE data for essentially subjective forecasting has been discussed and demonstrated in Chapters II through V. Many attempts also have been made to use this information from a single station to objectively forecast certain meteorological factors, such as visibility or precipitation amounts. These efforts involve the use of statistical techniques, generally leading to multivariate regression or Markov chain relationships. A survey of statistical procedures for this purpose is provided by Whiton and Berecek (1982).

In general, such statistical relationships are tied closely to the climatology of the given station and frequently show only small improvement over climatology or persistence. Furthermore, the prediction relationships usually hold for only a given station or differ from one station to another, i.e., require different input data, for the same predictand. Nevertheless, the approach has value when information from one's own station is the only data available or for the new forecaster in the station who is not yet familiar with local conditions. Summaries of some of the types of studies that have been made for a variety of predictands are provided below.

A procedure for predicting the ceiling and visibility in five category groups at 22 aviation terminals in the United States has been developed by Crisci (1973) and Crisci and Lewis (1973). Forecasts can be made at 3 h intervals from 4 to 16 h into the future. A similar study was performed by Allen (1970) for four stations having widely different climatological regimes. Allen also stratified his input data according to season. Testing of his relationships for independent data showed an improvement over climatology of 15%.

Druyan (1982) describes a technique for forecasting the 12 h accumulated rainfall during the rainy season at Bet Degan, Israel. The relationship used is based upon the mixing ratio values from the surface to 600 mb. A test of the relationship showed that it fell far short of predicting the correct amount of precipitation when the rain episode produced in excess of 10 mm of rain. A separate equation, but also depending upon the mixing ratio values, was developed to discriminate between rainfall amounts of $\leq$ 2 mm from amounts >2 mm. A

test of this equation showed that it had good ability to distinguish between rain/no rain situations. The Druyan method is useable with best results with a moist layer below and dry air aloft. A Markov chain model was used by Miller and Leslie (1984) to predict a rain/no rain condition with cloud cover amounts as antecedent states for five cities in Australia. Summer and winter seasons were separated. Skill scores calculated for a test of this model using 10 years of independent data showed it to be a significant improvement over persistence and climatology. These authors cite a number of other references in which the Markov Chain model has been used.

Examples of attempts to forecast summertime stratus along the California coast can be found in the work of Sweet (1983) and Grover and Jacobs (1983) using regression relationships. Ball et al. (1965) have developed a method which does not provide a prediction, but instead gives a diagnosis of the dew point spread at the mandatory pressure levels from 850 to 400 mb. The method involves the use of a decision tree using surface weather observation elements. Additional information on this procedure is given in the Appendix, Part C.

Miller (1981) has developed a statistical procedure called GEM for generalized equivalent (or exponential) Markov. This scheme uses only one local surface observation as a predictor and provides forecasts of the same weather elements at 1 h increments out to 12 h. The basis for the technique is multivariate regression which was used to establish the coefficients of forecast equations with input of nearly four million observations from 41 stations spread over the United States. As a consequence, there is a generalized climatology built into the system. The coefficients for the regression equations are kept constant as new forecast values are calculated, iteratively, using the previous hour's predicted values as predictors for the next hour. Miller states that for a reduced field of predictands, the calculations can be made on a hand-held calculator. Upon testing his scheme, Miller found that he obtained better Brier scores than did persistence for the 24 weather elements in the prediction and that improved skill was obtained when station-adjusted climatology was introduced into the prediction equations.

The reader also may wish to examine work by Russo et al. (1964) for a procedure for prediction of winds aloft. Objective statistical methods for forecasting surface temperature are provided by Klein (1966) and Hansen and Driscoll (1977). Finally, Williford et al. (1974) have developed a method for the objective prediction of thunderstorm activity.

PART B PRACTICAL OBSERVATIONS

1. Operating Station Suffers Communication Outage. In a case where an operating station does not receive outside data, many forecasting aids for the station are useable.

   a. Persistence-Probability Tables. The persistence-probability tables have been computed for USAF/AWS stations and are easy to use as a guide. For the background and methods used see McCabe (1968).

   b. Statistical Forecasts. Regression equations and Markov chain equations can be computed for an operating station that has at least 10 y of data. ETAC has already prepared many of these and they are available.

   c. Objective Forecasts. The Detachment Terminal Reference File has objective forecast procedures and "rules-of-thumb" which apply to that station. Many "rules-of-thumb" apply to only one station or area and they should be reviewed before trying them in a different area.

2. Statistical Forecast at a New Station. A forecaster moving into an area where the climatic data has not been processed into the required equations will not be able to use these methods. During the last several years many "single station forecast" methods or studies have been published in the literature and some of it was referenced in PART A of this chapter. However, all of them require the climatic background or synoptic data from a base file. A forecaster in a "new" area will not have the necessary background input. He will have to use his knowledge of what usually happens (climatology), keep watching the celestrial dome, and make a subjective forecast of changes from persistence, using the models.

REFERENCES

Air Weather Service, 1960: Use of asynoptic data in analysis and forecasting. TR 225, USAF Air Wea. Ser., 27 pp.

———, 1979a: Meteorological techniques. AWSP 105-56, USAF Air Wea. Ser., 181 pp.

———, 1979b: Use of the skew T-log p diagram in analysis and forecasting. TR 79/006, USAF Air Wea. Ser., 142 pp.

———, 1983: Meteorological Codes. AWSR 105-24, USAF Air Wea. Ser., 205 pp.

Allen R. A., 1970: Single-station prediction of ceiling and visibility. FAA-RD-70-26, 51 pp.

Anthes, R. A., 1982: Tropical cyclones, their evolution, structure and effects. Met Monograph #19 AMS, 208 pp.

Atkinson, G. D., 1971: Forecasters' guide to tropical meteorology. TR 240, USAF Air Wea. Ser., 348 pp.

Ball, J. T., D. A. Chisholm, P. V. Luty and K. W. Veigas, 1965: The diagnosis of upper-air humidity from surface observations. Tech. Rep. ESD-TR-65-1, The Travelers' Research Center, Inc., AF 19 (628)-3437, 195 pp.

Berry, F. A. and E. Bollay, 1945: Air masses. Handbook of Meteorology, McGraw-Hill Book Company, Inc., 604-637.

Boyden, M., V. Oliver and J. Holland, 1942: An example of a field forecast. Miscellaneous Rep. #6, University of Chicago Institute of Meteorology, 13-28.

Brundidge, K. C., 1965: The wind and temperature structure of nocturnal cold fronts in the first 1,420 feet. Mon. Wea. Rev., 93, 587-603.

Chary, H. A. and R. L. Lininger, 1979: A method of determining winds aloft. AWSP 105-56(C2), USAF Air Wea. Ser., 8-1.2-1 to 8-1.2-6.

Crisci, R. L., 1973: Automated ceiling and visibility forecasts -- an evaluation of an operational test. FAA-RD-73-182, 18 pp.

——— and F. Lewis, 1973: Automated probability forecasts of ceiling and visibility based on single-station data. FAA-RD-73-13, 68 pp.

Davenport, A. G., 1961: The application of statistical concepts to the wind loading of structures. Proc. Inst. Civ. Engrs., 19, 6480.

Druyan, L. M., 1982: Automated precipitation forecasts using a single-station radiosonde profile. Bull. Amer. Meteor. Soc., 63, 499-502.

Duffield, G. F. and G. D. Nastrom, 1984: Equations and algorithms for meteorological applications in Air Weather Service. TR 83/001, USAF Air Wea. Ser., 58 pp.

Dwyer, L. J., 1960: Antarctic Meteorology, Pergamon Press, 483 pp.

Federal Meteorological Handbook No. 1 as revised. US DOC, US DOD, US DOT.

Gavrilova, M. K., 1966: Radiation Climate of the Arctic. Clearinghouse 1166 5 1058, 178 pp.

Grover, 1979: Stability indices. AWSP 105-56, USAF Air Wea. Ser., 3-1.5-1 to 3-1.5-2.

Grover, D. P. and P. A. Jacobs, 1983: Low-level stratus prediction using statistical regression. Naval Postgraduate School Report 55-83-034, 77 pp.

Hansen, J. E. and D. M. Driscoll, 1977: A mathematical model for the generation of hourly temperatures. J. Appl. Meteor., 16, 935-948.

Henry, W. K., 1965: The ice jam floods of the Yukon River. Weatherwise, 18, 80-85.

———, J. F. Griffiths and F. J. Klein Jr., 1972: Research on tropical rainfall patterns and associated mesoscale systems. T. R. ECOM-0123-2, Dept. of Meteor. Project 778 72-3-T, 1-12.

———, 1974: The tropical rainstorm. Mon. Wea. Rev., 102, 717-725.

Holton, J. R., 1979: An Introduction to Dynamic Meteorology. Academic Press, 391 pp.

Hoxit, L. R., 1973: Variability of boundary layer winds. Rep. # CSU-ATSP-199, NSF GA-32589X1, Atmospheric Sciences Dept., Colorado State University, 157 pp.

Jenkins, G. R., 1945: Diurnal variation of the meteorological elements. Handbook of Meteorology, McGraw-Hill Book Company, Inc., 746-753.

Jones, W. W., 1943: Weather-map construction and forecasting in the westerlies from single-station aerological data. Miscellaneous Rep. #7, University of Chicago Institute of Meteorology, 83 pp.

Klein, W. H., 1966: Objective forecasts of surface temperature from one to three days in advance. J. App. Meteor., 5, 137-147.

Krishnamurti, T. N., 1985: Summer monsoon experiment -- a review. Mon. Wea. Rev. 113, 1590-1626

McCabe, J. T., 1968: Estimating conditional probability and persistence. TR 208, USAF Air Sea. Ser., 20 pp.

Miller, A. J. and L. M. Leslie, 1984: Short-term single-station forecasting of precipitation. Mon. Wea. Rev., 112,1198-1205.

Miller, R. C., 1972: Notes on analysis and severe-storm forecasting procedures of the military weather warning center. TR 200 Rev. w/ change, 1975, USAF Air Wea. Ser., 171 pp.

Miller, R. G., 1981: GEM: a statistical weather forecast procedure. NOAA Technical Rep. NWS 28, 103 pp.

Moreland, W. B., 1965: Estimating meteorological effects on radar propagation. TR 183, Vol I, USAF Air Wea. Ser., 198 pp.

Oliver, V. J. and M. B. Oliver, 1945: Weather analysis from single-station data. Handbook of Meteorology, McGraw-Hill Book Company, Inc., 858-879.

Palmer, C. E., C. W. Wise, L. J. Stempson and G. H. Dunkan, 1955: The practical aspect of tropical meteorology. TR 241 Vol I, USAF Air Wea. Ser., 195 pp., Vol II (1961) 180 pp.

Petterssen, S., W. C. Jacobs and B. C. Haynes, 1956: Meteorology of the arctic. US Navy OPNAV P03-3, 207 pp.

Reed, J. W., 1967: Some notes on forecasting of winds aloft by statistical methods. J. Appl. Meteor., 6, 360-372.

Reed, R. J. et al., 1956-1959: Arctic meteorology -- a series of reports. Dept. of Meteor., University of Washington.

Riehl, H., 1979: Climate and Weather in the Tropics. Academic Press, 609 pp.

Rossby, C. G., V. Oliver and M. Boyden, 1942: Weather estimates from local aerological data. Miscellaneous Rep. #2, University of Chicago Institute of Meteorology, 47 pp.

Rubin, J. J. (Ed.), 1966: Studies in Antarctic Meteorology. Am. Geop. Union, 231 pp.

Russo, J. A., I. Enger and E. Sorenson, 1964: A statistical approach to the short-period prediction of surface winds. J. Appl. Meteor., 3, 126-131.

Sadler, J. C., 1976a: A role of the tropical upper tropospheric trough in early season typhoon development. Mon. Wea. Rev. 104, 1266-1278.

———, 1976b: Tropical cyclone initiation by the tropical upper tropospheric trough. NAVENVPREDRSCHFAC Lecture Paper. No. 2-76, 103 pp.

———, 1978: Mid-season typhoon development and intensity changes and the tropical upper tropospheric trough. Mon. Wea. Rev. 106, 1137-1152.

Shaw, D. B., 1978: Meteorology over the tropical oceans. Royal Meteor. Soc., 278 pp.

Simpson, R. H. and H. Riehl, 1981: The Hurricane and its Impact. Louisiana State Univ. Press, 398 pp.

Slade, D. H. (Ed.), 1968: Meteorology and atomic energy. United States Atomic Energy Commission, Office of Information Services, 445 pp.

Sutcliffe, R. C., 1958: Polar Atmosphere Symposium, Part I. Pergamon Press, 341 pp.

Sweet, W. A., 1983: Short-range local forecasting of coastal summer stratus by statistical techniques. NAVENVPREPSCHFAC TR 83-06, US Navy, 52 pp.

Weikes, K., 1957: Polar Atmosphere Symposium, Part I. Pergamon Press, 212 pp.

Weller, G. and S. A. Bowling, 1973: Climate of the Arctic. Geophysical Institute, University of Alaska, 435 pp.

Whiton, R. C. and E. M. Berecek, 1982: Basic techniques in environmental simulation. USAETAC/TN-82/004, 144 pp.

Williford, W. O., M. C. Carter and P. Hsieh, 1974: A Bayesian analysis of two probability models describing thunderstorm activity at Cape Kennedy, Florida. J. Appl. Meteor., 13, 718-725.

Winn-Nielsen, A., 1979: Compendium of Meteorology Vol II. WMO HO 364, 428 pp.

Zandgren, G. L. and G. M. Leies, 1945: Pressure tendency as an aid in single station forecasting for tropical areas. TR 105-58, USAF Air Wea. Ser., 14 pp.

INDEX OF THE APPENDIX

APPENDIX

PART A

List of materials and equipment for doing single station analysis.

Copy or extracts of this manual

Time cross-section forms

Hodographs (A Dead Reckoning Computer type E-6B or equivalent can be used in place of the hodograph. This computer does have conversion scales and the standard atmosphere information in easy-to-use slide rule format.)

Maps of the area

Straight edge

A scale with units that match the hodograph is handy

Triangles

Plotting pens

Pencils, assorted colors

Skew T-Log P Diagrams

Calculator; Many calculators can be programmed prior to starting single station analysis procedures.

Tables of geostrophic wind speed vs isobar or contour spacing, or other charts, such as the Bellamy gradient wind scale

Compass and/or dividers

APPENDIX

PART B

List of Symbols and Abbreviations.

List of Symbols

A_a	area of vertical wall of triangle
A_i	area of i^{th} face
A'	area of triangle
α	wind direction (from)
B (subscript)	bottom of layer
ε	orientation angle of triangle legs
$\vec{C}$	the three dimensional velocity vector
° C	Celsius temperature scale
cm	centimeter
D	day
D_1	first day, D_2 second day, etc.
∇	the del operator
Δ	difference or change of ___
Δ	determinent of the coefficients of the homogeneous form of IV-6
E	east
ENE	east-northeast

List of Abbreviations

ACSL N	Altocumulus Standing Lenticularis are north of the station
AGL	Above Ground Level
AWSM	Air Weather Service Manual
AWSP	Air Weather Service Pamphlet
AWSTR	Air Weather Service Technical Report
BKN	Broken clouds 6/10 to 9/10
CAT	Clear Air Turbulence
CLR	Clear Sky
cA	Continental Arctic
cP	Continental Polar
cPk	Continental Polar, cold
cT	Continental Tropic
ETAC	Environmental Technical Applications Center
E	Equatorial

°F	Fahrenheit temperature scale
f	Coriolis parameter = $2\Omega \sin\phi$
ft	feet
g	acceleration due to gravity
Γ	lapse rate (temperature)
Γ_d	dry adiabatic lapse rate (temperature)
GATE	Global Atmospheric Research Program Atlantic Tropical Experiment
GLW	Gradient Level Wind
GMT	Greenwich Mean Time
gpm	Geopotential Meters
H	height of pressure levels
HHH	height of pressure levels from a RAOB code
h	hour
$\hat{i}$	unit vector in the x coordinate direction
$\hat{j}$	unit vector in the y coordinate direction
°K	Kelvin temperature scale
$\hat{k}$	unit vector in the z coordinate direction
km	kilometers
kt	knots
KI	"K" Index
LI	Lifted Index
LST	Local Standard Time
Lat	Latitude
mb	millibar
mi	mile
mm	millimeter
MVG	Moving
MSL	Mean Sea Level
mP	Maritime Polar
mT	Maritime Tropical

N	north	NATO	North Atlantic Treaty Organization
NE	northeast		
n	horizontal distance		
n mi	nautical mile		
$\hat{n}$	unit vector normal to a boundary		
o	as subscript refers to conditions at the centroid of triangle, or refers to value at initial time	OVC	Overcast 10/10
Ω	magnitude of earth's angular velocity		
p	pressure (mb)	PA	Pressure Altitude
p_1, p_2, ...	identified pressures	PIREP	Pilot Report Code
ϕ	latitude		
Q	any quantity which varies in space and time		
R	radius of curvature	RAOB	Radiosonde Observation
RH	relative humidity	RAWIN	Radiosonde Observation <u>and</u> winds aloft
R_d	gas constant for dry air	RECCO	Weather Reconnaissance Code
r	mixing ratio		
$\bar{r}$	mean mixing ratio		
S	south	SCTD	Scattered Clouds 1/10 to 5/10
SW	southwest	SI	Showalter Index
s	distance -- side of triangle	SSA	Single Station Analysis
$d\sigma$	element of area	SWEAT	Sweat Index

T	temperature
T (subscript)	top of layer
$\bar{T}$	mean temperature
T_c	convective temperature
T_d	dew-point temperature
T_v	virtual temperature
$\bar{T}_v$	mean virtual temperature
T_{500}	temperature at 500 mb
T'	lifted parcel's temperature at 500 mb
t	time
t (subscript)	at time t
θ	potential temperature
θ_E	equivalent potential temperature
TTI	Total-Totals Index
TUTT	Tropical Upper Troposphere Trough
u	west wind component
$\bar{u}$	average west wind component
USAF/AWS	United States Air Force/ Air Weather Service
V	wind speed
$\bar{V}$	average wind speed
$\vec{V}$	wind vector
V_g	geostrophic wind speed
$\vec{V}_g$	geostrophic wind vector
V_{gr}	gradient wind
V_n	component of wind normal to a lateral boundary
$\bar{V}_n$	average component of the wind normal to the surface
$\vec{V}_n$	wind vector normal to a surface
V_o	wind speed at centroid of triangle
V_t	thermal wind speed
$\vec{V}_t$	thermal wind vector
$\vec{V}_1$, $\vec{V}_2$	identifier wind vector
v	north wind component
$\bar{v}$	average north wind component

W	west
$\overline{w}$	average vertical motion
x	horizontal distance east-west
y	horizontal distance north-south
z	height
ζ_A	absolute vorticity
ζ_o	relative vorticity

z-t Diagram a chart with z (height) as the ordinate and t (time) as the abscissa

APPENDIX

PART C

Relative Humidity in percent for different temperatures ° C,

dew-point spread ° C

and a $T-T_d$ ° C vs cloud amount "Rule of Thumb".

a. 1000 mb

T	$T-T_d$	0	2	4	6	8
30		100	88.4	78.3	69.2	61.2
20		100	87.3	76.7	68.0	58.7
10		100	87.1	76.0	66.2	57.1
0		100	86.9	74.8	64.0	55.1
-10		100	84.9	72.1	61.5	51.9
-20		100	84.6	71.2	59.0	48.7

b. 850 mb

T	$T-T_d$	0	2	4	6	8
25		100	88.3	78.1	68.2	60.2
15		100	87.6	76.6	66.6	57.9
5		100	86.6	75.2	64.7	56.7
-5		100	85.8	73.3	62.8	52.8
-15		100	84.3	71.1	60.4	50.3

c. 700 mb

T	$T-T_d$	0	2	4	6	8
15		100	87.7	76.8	67.1	58.1
5		100	86.5	75.5	65.1	56.3
-5		100	85.9	74.2	63.3	53.5
-15		100	84.1	71.8	60.6	50.6
-25		100	81.1	67.6	56.8	45.9

d. 500 mb

T	$T-T_d$ 0	2	4	6	8
-10	100	85.8	72.4	61.6	51.8
-20	100	84.1	70.7	58.6	48.4
-30	100	82.8	68.8	55.5	45.3
-40	100	--	--	--	--

Relationship of dew-point spread in ° C and cloud amounts, an old "Rule of Thumb".

$T-T_d$	CLOUD AMOUNT
0-2	OVERCAST
2-4	BROKEN
4-6	SCATTERED
6+	CLEAR

APPENDIX

PART D

Estimating relative humidity aloft.

After Ball, et al. (1965)

The moisture aloft may be estimated when radiosonde data are not available. Ball, et al. 1965 suggest a procedure which only a part has been extracted for use in this manual. The procedure is to take a series of surface observations, then follow through a decision tree to an answer, expressed in $T\text{-}T_d$. The decision trees are made for different pressure levels. Enter the tree at the upper left hand block. If one of the ww types is occurring go right on the YES line. If not, go down on the NO line. Repeat until a $T\text{-}T_d$ is determined or the Statistical Approach block is reached. In the field without instruments the Statistical Approach is not practical, and estimate $T\text{-}T_d$ to be greater than 5° C. The decision trees for four levels follow the tables of ww, w, and cloud types.

After a $T\text{-}T_d$ has been determined the forecaster may wish to check the "Rule of Thumb" in the Appendix Part C.

ww

21 Rain (NOT freezing and NOT falling as showers) during past hour, but NOT at time of observation.

22 Snow (NOT falling as showers) during past hour, but NOT at time of observation.

25 Showers of rain during past hour but NOT at time of observation.

26 Showers of snow, or of rain and snow, during past hour, but NOT at time of observation.

50 Intermittent drizzle (NOT freezing) slight at time of observation.

51 Continuous drizzle (NOT freezing) slight at time of observation.

52 Intermittent drizzle (NOT freezing) moderate at time of observation.

53 Continuous drizzle (NOT freezing) moderate at time of observation.

54 Intermittent drizzle (NOT freezing) thick at time of observation.

55 Continuous drizzle (NOT freezing) thick at time of observation.

56 Slight freezing drizzle.

57 Moderate or thick freezing drizzle.

58 Fog, depositing rime, sky discernible.

59 Fog, depositing rime, sky NOT discernible.

60 Intermittent rain (NOT freezing) slight at time of observation.

61 Continuous rain (NOT freezing) slight at time of observation.

62 Intermittent rain (NOT freezing) moderate at time of observation.

63 Continuous rain (NOT freezing) moderate at time of observation.

64 Intermittent rain (NOT freezing) heavy at time of observation.

65 Continuous rain (NOT freezing) heavy at time of observation.

66 Slight freezing rain.

67 Moderate or heavy freezing rain.

68 Rain or drizzle and snow, slight.

69 Rain or drizzle and snow, moderate or heavy.

70 Intermittent fall of snowflakes, slight at time of observation.

71 Continuous fall of snowflakes, slight at time of observation.

72 Intermittent fall of snowflakes, moderate at time of observation.

73 Continuous fall of snowflakes, moderate at time of observation.

74 Intermittent fall of snowflakes, heavy at time of observation.

75 Continuous fall of snowflakes, heavy at time of observation.

79 Ice pellets (sleet, U.S. definition)

80 Slight rain shower(s).

81 Moderate or heavy rain shower(s).

82 Violent rain shower(s).

83 Slight shower(s) of rain and snow mixed.

84 Moderate or heavy shower(s) of rain and snow mixed.

85 Slight snow shower(s).

86 Moderate or heavy snow shower(s).

87 Slight shower(s) of soft or small hail with or without rain or rain and snow mixed.

88 Moderate or heavy shower(s) of soft or small hail with or without rain or rain and snow mixed.

89 Slight shower(s) of hail, with or without rain or rain and snow mixed, not associated with thunder.

$\underline{W}$

0 Clear or few clouds.

6 Rain.

7 Snow, or rain and snow mixed, or ice pellets (sleet).

8 Shower(s).

9 Thunderstorm, with or without precipitation.

$\underline{C_L}$

0 None.

2 Cu of considerable development, generally towering, with or without other Cu or Sc bases all at same level.

3 Cb with tops lacking clear-cut outlines, but distinctly not cirriform or anvil-shaped; with or without Cu, Sc, or St.

4 Sc formed by spreading out of Cu; Cu often present also.

7 Fs and/or Fc of bad weather (scud).

8 Cu and Sc (not formed by spreading out of Cu) with bases at different levels.

9 Cb having a clearly fibrous (cirriform) top, often anvil-shaped, with or without Cu, Sc, St, or scud.

$\underline{C_M}$

1 Thin As (most of cloud layer semitransparent).

2 Thick As, greater part sufficiently dense to hide sun (or moon), or Ns.

7 Double-layered Ac, or a thick layer of Ac, not increasing; or Ac with As and/or Ns.

$\underline{N_h}$

0 No clouds.

1 Less than one-tenth or one-tenth.

2 Two and three-tenths.

3 Four-tenths.

4 Five-tenths.

5 Six-tenths.

6 Seven and eight-tenths.

7 Nine-tenths or overcast with openings.

8 Completely overcast.

9 Sky obscured.

N_T Amount of sky cover, clear, scattered, broken, and overcast.

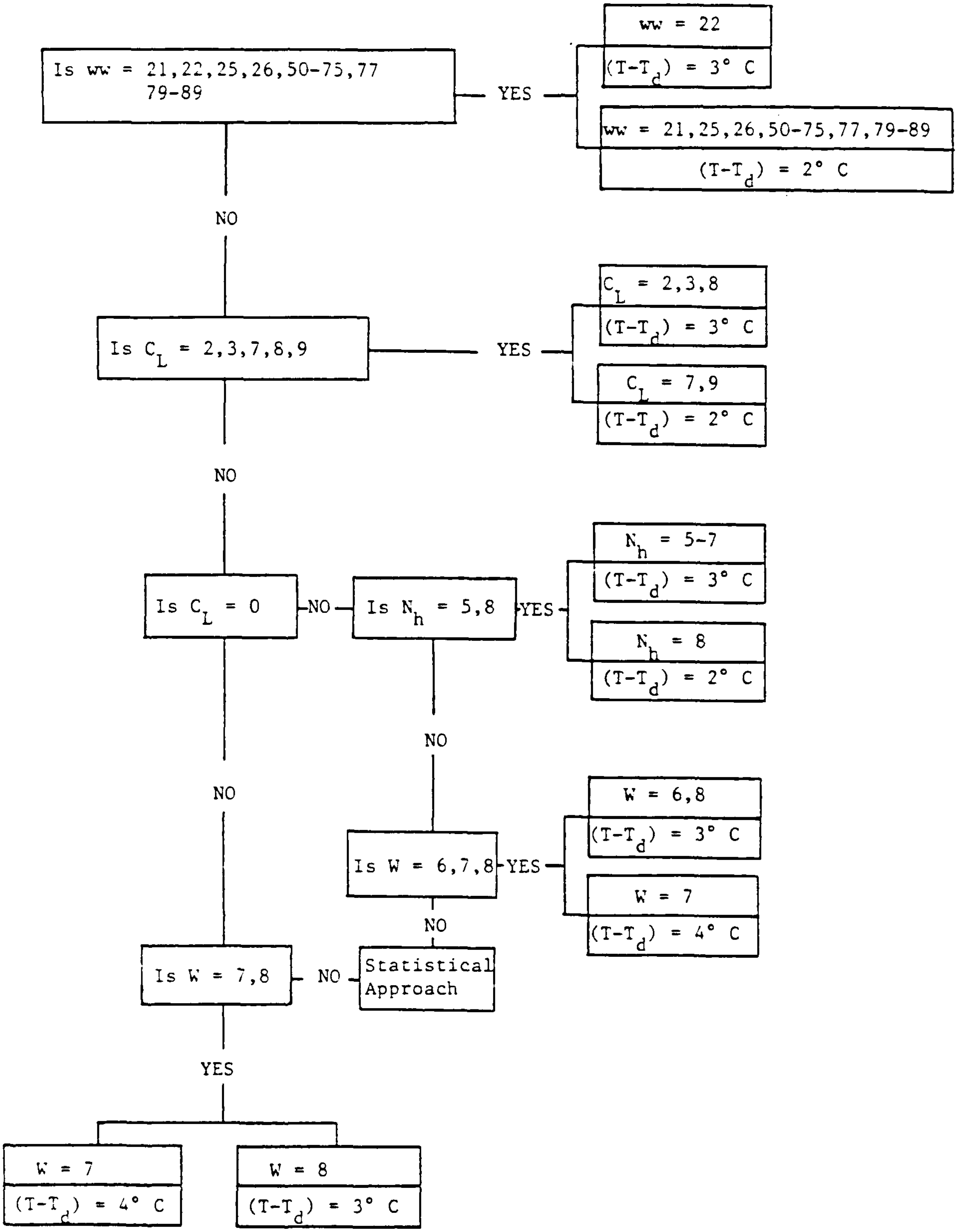


FIG. D-1 Recommended 850 mb decision tree.

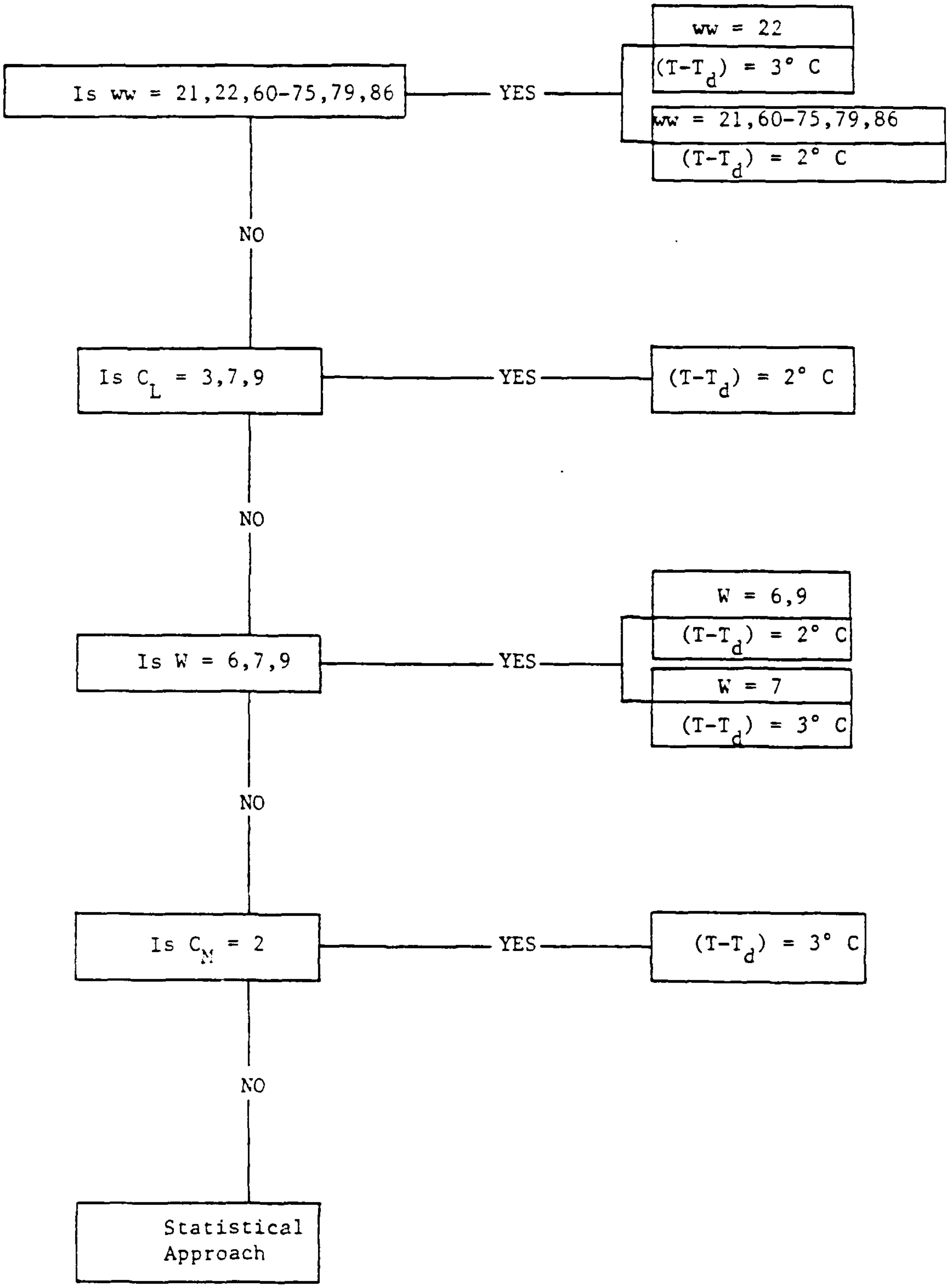


FIG. D-2 Recommended 700 mb decision tree.

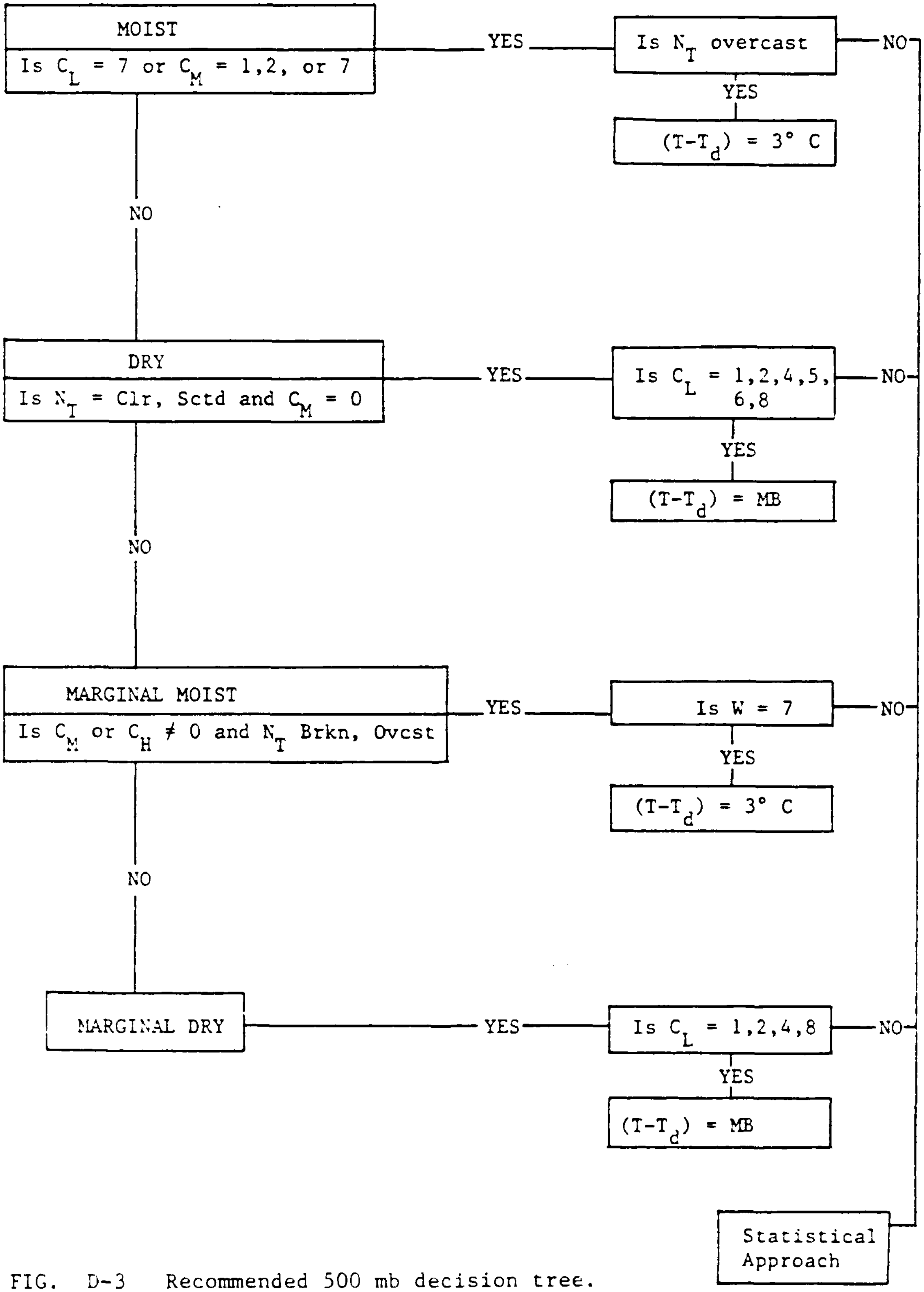


FIG. D-3 Recommended 500 mb decision tree.

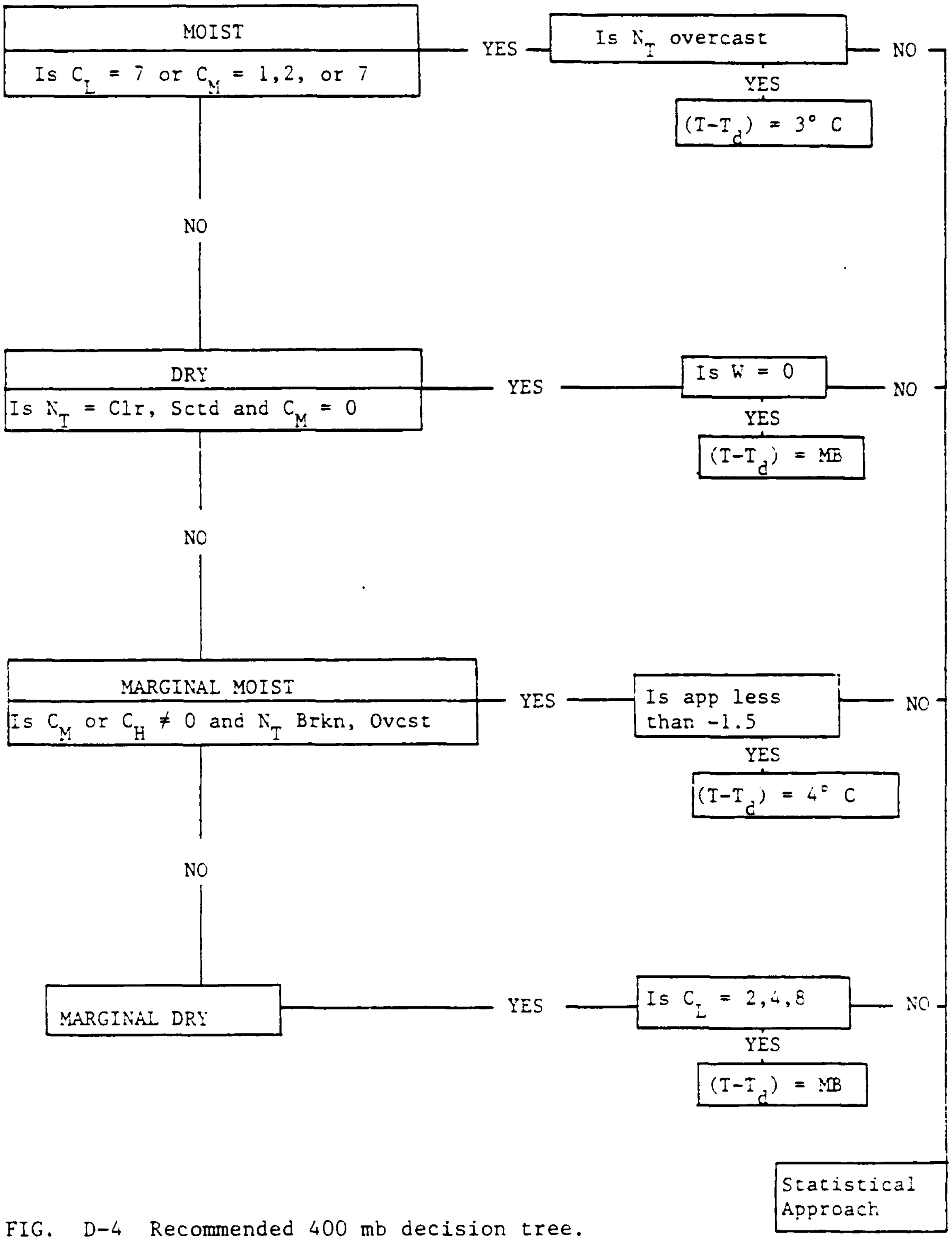


FIG. D-4 Recommended 400 mb decision tree.

APPENDIX

PART E

Beaufort wind scale.

At times an observer may need to report wind speed without the aid of an anemometer. At such a time he may estimate the force of the wind according to the descriptive terms in the Beaufort wind table. The speeds indicated, it should be remembered, are for a level about 33 feet above level ground. It should also be remembered that the Beaufort scale does not incorporate a correction for turbulence, or gustiness. If the observer does not take gustiness into account when it exists, than in using the movement of trees, etc., as a measure of wind force, it is possible that his estimated value of force, and hence of speed, will be too high.

Wind-Speed Equivalents

Descriptive Word	Velocity, Knots	Specifications for Estimating Velocities
Calm	Less than 1	Smoke rises vertically
	1 to 3	Direction of wind shown by smoke drift but not by wind vanes.
Light	4 to 6	Wind felt on face; leaves rustle; ordinary vane moved by wind.
Gentle	7 to 10	Leaves and small twigs in constant motion; wind extends light flag.
Moderate	11 to 16	Raises dust and loose paper; small branches are moved.
Fresh	17 to 21	Small trees in leaf begin sway; crested wavelets form on inland waters.
	22 to 27	Large branches in motion; whistling heard in telegraph wires; umbrellas used with difficulty.
Strong	28 to 33	Whole trees in motion; inconvenience felt in walking against the wind.
	34 to 40	Breaks twigs off trees; generally impedes progress.
Gale	41 to 47	Slight structural damage occurs (chimney pots and slate removed).
	48 to 55	Trees uprooted; considerable structural damage occurs.
Whole gale	56 to 65	Rarely experienced; accompanied by widespread damage.
Hurricane	Above 65	

APPENDIX

PART F

Determining directions and winds aloft.

DIRECTIONS. Wind has to include both direction and speed. To determine the wind direction the observer must be orientated with the cardinal directions. A general first start is the direction of sun or moon rise and set. However at different times of the year the direction of rise may be 30 or 40 degrees from east and west. One of the best methods is to take a stick (3 or 4 ft long) or narrow board and put one end in the ground and have it nearly vertical by using a plumbline (rock tied to a string). As noon approaches make marks on the ground (small pegs) at time intervals of about 30 minutes of the location of the shadow of the top of the stick. The marks will be in an almost straight line which will be east-west. North and south may now be determined.

UPPER LEVEL WINDS FROM CLOUDS. The direction of winds at various levels may be determined by observing the movement of clouds. The height of the cloud must be estimated and then its movement may be determined. A first approximation may be made by watching a cloud blow away from you. An improvement is to use a stick, board or piece of cord about one meter long. Lay this on the ground and keep turning it until it is parallel to the cloud movement. The method is applied best with a reflector under the stick so the cloud may be seen in the reflection.

In Fig. F-1 the observer sights a cloud across the top of a tree or flagpole. He moves from a to b so as to keep the cloud centered in his view across the fixed point. His motion from a to b will then be in the direction opposite to that of the cloud (A to B). If the observer moves in some other direction, he will not be able to keep the cloud centered on the fixed point. This method may be applied to a cloud at a distance or one overhead but does not provide the speed. A somewhat similiar but more sophisticated technique using a clinometer is described by Chary and Lininger (1979); it provides both direction and speed.

A simple method which provides both direction and speed for clouds overhead is illustrated in Fig. F-2. The observer holds a ruler at arm's length overhead in alignment with the motion of the clouds. The distance d is the observed displacement of a cloud relative to the ruler while D is

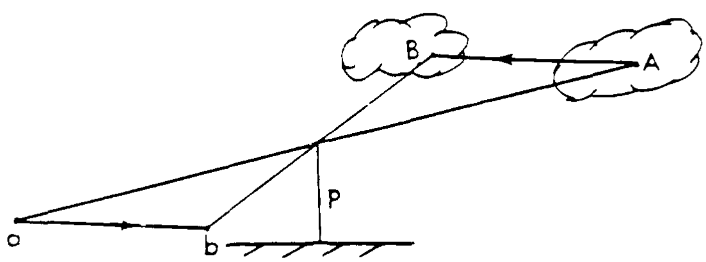


FIG. F-1 Finding cloud direction. Using pole P the observer starts at (a) and moves to (b) as the cloud moves from (A) to (B). The direction ba is the direction the cloud is moving. The observer keeps the cloud in line with the top of the pole.

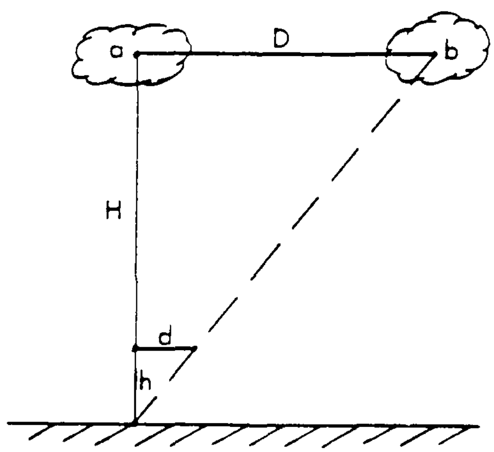


FIG. F-2 Estimating speed of clouds. The observer holds a ruler overhead and watches the cloud move from (a) to (b). H represents the height of the cloud. D is the distance the cloud moves. h is the distance the ruler is above the eye and d is the distance indicated on the ruler.

its displacement in space. The estimated height of the cloud, H, is the distance from the ground to a and the length of the observer's arm is h. Then from the law of similiar triangles

$$\frac{D}{d} = \frac{H}{h} \text{ or } D = \frac{Hd}{h} .$$

If the time of the displacement is t, then D = Vt and

$$V = \frac{H \text{ X } d}{h \text{ X } t} .$$

A better arrangement would be to have a spoked wheel hung about 6 ft above the observer's head. Then the radius of the wheel would be known, and the direction determined by the spokes. This provides a more stable platform than holding a ruler.

APPENDIX

PART G

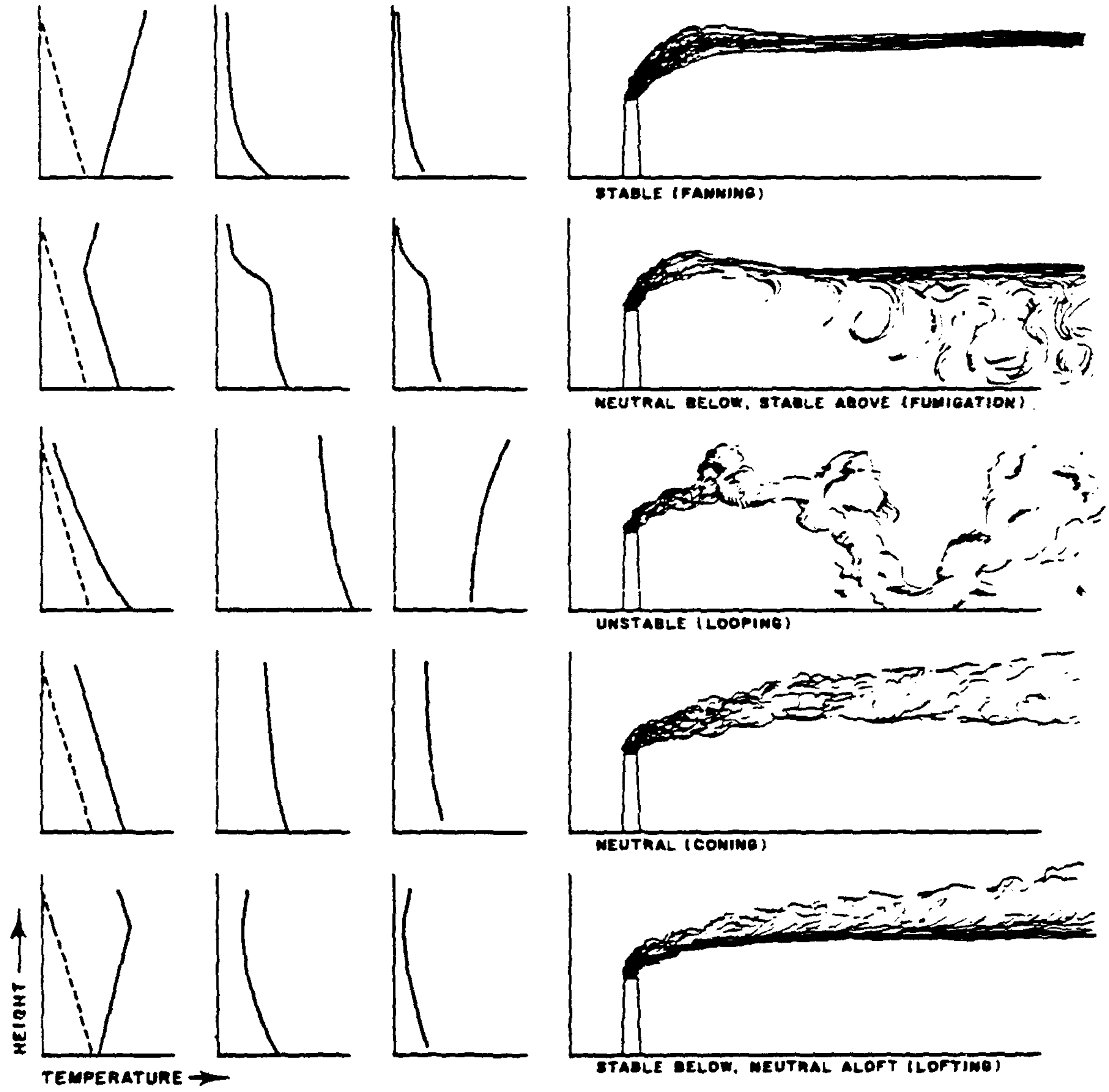


FIG. G-1 Low level lapse rates, wind, and vertical diffusion.
After Slade (1968)

The dashed curves in the left-hand column of diagrams show the adiabatic lapse rate, and the solid lines are observed temperature lapse rates. The middle column shows the variation with height of the variation of the standard deviation of the wind direction with height. The right-hand column shows the variation of the standard deviation of the wind speed with height.

a. Fanning. This condition occurs with stable conditions. The low-frequency components of the turbulence are suppressed, especially in the vertical, so the vertical spread of the plume is very small. The plume may show a slow meandering of direction. With respect to a long-term mean wind, this meandering produces an effective spreading of the plume which is much greater laterally than vertically.

b. Fumigation. There is a brief period after sunrise when the lapse rate is stable aloft but unstable near the ground. When the top of the unstable layer reaches the height of the stack, the diffusion toward the ground can be large while the upward diffusion remains small. A high concentration of pollutant can then suddenly be brought to ground level.

c. Looping illustrates the much larger scale of the turbulent eddies under unstable conditions. Diffusion is very rapid so that points away from the axis of the plume, but near the source, experience a greater concentration than under more stable conditions. On the other hand, further downwind, concentrations are smaller than with more stable lapse rates.

d. Coning is characteristic of nearly equal intensities of turbulence in the vertical and horizontal directions. This is characteristic of temperature gradients between isothermal and adiabatic. The scale of the eddies, or the low-frequency energy of the turbulence, is seen to be less than experienced with looping, unstable conditions, but more than in the stable conditions which produce fanning. Coning is characteristic of strong winds and cloudy skies.

e. Lofting is the opposite of fumigation. When an inversion is present below the stack, but not above, the preferred direction of the diffusion is upward. This situation is very desirable from the point of view of avoiding hazardous exposure at the ground. Of course, it applies only to elevated sources of contaminant.

APPENDIX

PART H

Climatic Data for Station #3 for February.

Mean Daily Maximum Temperature	42.7° F
Mean Daily Minimum Temperature	24.5° F
Mean Monthly Temperature	33.6° F
Extreme Maximum Temperature	82° F
Extreme Minimum Temperature	-6° F
Normal Precipitation	1.72 inches water equivalent
Maximum Monthly Precipitation	2.96 inches water equivalent
Minimum Monthly Precipitation	0.31 inches water equivalent
Maximum Precipitation in 24 h	1.33 inches water equivalent
Maximum Monthly Snow	16.1 inches
Maximum 24 h Snow	11.8 inches
Percent of time ceiling < 500 ft and/or visibility < 1 mile	5%
Percent of time ceiling is 500' to 900' with visibility ≥ 1 mile and/or visibility 1 to 1½ miles and ceiling ≥ 500'	11%
Percent of time ceiling ≥ 100' with visibility ≥ 3 miles	81%
Hours per month wind exceeds 25 kt	6 h
Percentage of occurrence of 6/10 or more cloud cover surface to 5000 ft MSL	35%
Percentage occurrence of 6/10 or more cloud cover 5000' to 10000' MSL	16%
Percentage occurrence of 6/10 or more cloud cover 10000' to 15000' MSL	9%
Percentage occurrence of 1/10 or more cloud cover 15000' to 20000' MSL	6%

APPENDIX

PART 1

Table for the spacing in degrees of latitude of 2 mb surface isobars at geostrophic wind speeds at different latitudes. If wind is greater than 50 kt, divide wind speed by 2, find spacing, and divide by 2.

kt \ ° Lat	5	10	15	20	25	30	35	40	45	50	55	60	65	70	75	80	85
5	55.0	27.6	18.5	14.0	11.3	9.6	8.4	7.5	6.8	6.3	5.8	5.5	5.3	5.1	5.0	4.9	4.8
10	27.5	13.8	9.3	7.0	5.7	4.8	4.2	3.7	3.4	3.1	2.9	2.8	2.6	2.5	2.5	2.4	2.4
11	25.0	12.5	8.4	6.4	5.2	4.4	3.8	3.4	3.1	2.8	2.6	2.5	2.4	2.3	2.2	2.2	2.2
12	22.9	11.5	7.7	5.8	4.7	4.0	3.5	3.1	2.8	2.6	2.4	2.3	2.2	2.1	2.1	2.0	2.0
15	18.3	9.2	6.2	4.7	3.8	3.2	2.8	2.5	2.2	2.1	1.9	1.8	1.8	1.7	1.6	1.6	1.6
20	13.7	6.9	4.6	3.5	2.8	2.4	2.1	1.9	1.7	1.6	1.5	1.4	1.3	1.3	1.2	1.2	1.2
25	11.0	5.5	3.7	2.8	2.3	1.9	1.7	1.5	1.4	1.2	1.2	1.1	1.0	1.0	1.0	1.0	1.0
30	9.2	4.6	3.1	2.3	1.9	1.6	1.4	1.2	1.1	1.0	1.0	0.9	0.9	0.8	0.8	0.8	0.8
35	7.8	3.9	2.6	2.0	1.6	1.4	1.2	1.1	1.0	0.9	0.8	0.8	0.8	0.7	0.7	0.7	0.7
40	6.9	3.4	2.3	1.8	1.4	1.2	1.0	0.9	0.8	0.8	0.7	0.7	0.7	0.6	0.6	0.6	0.6
45	6.1	3.1	2.0	1.6	1.2	1.1	0.9	0.8	0.8	0.7	0.6	0.6	0.6	0.6	0.6	0.5	0.5
50	5.5	2.8	1.8	1.4	1.1	1.0	0.8	0.7	0.7	0.6	0.6	0.6	0.5	0.5	0.5	0.5	0.5

APPENDIX

PART J

Table of the spacing in degrees latitude of 30 m contours at geostrophic wind speeds at different latitudes.

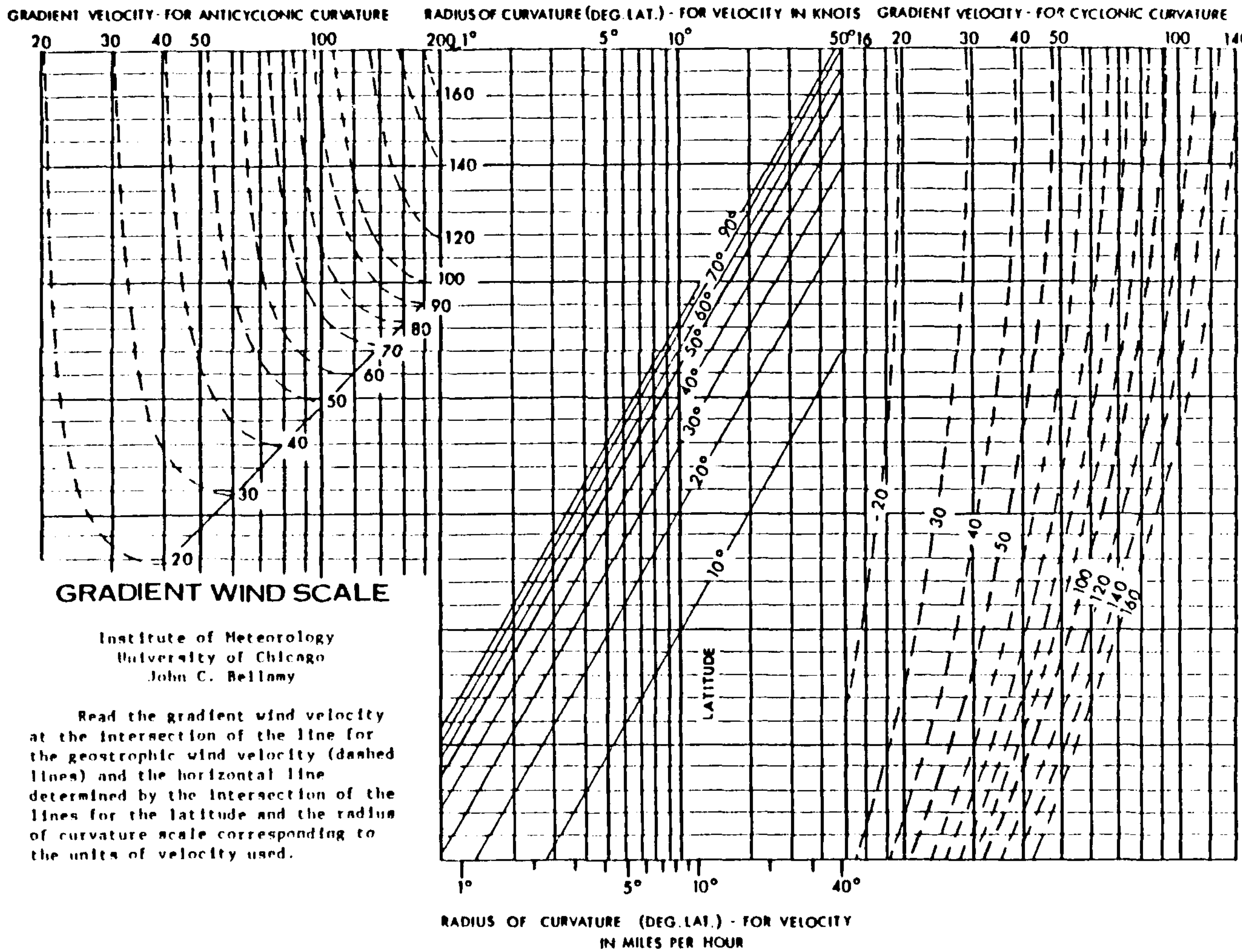
When wind speeds are greater than 100 kt, divide wind by 2, read latitude spacing and divide by 2.

kt \ ° Lat	5	10	15	20	25	30	35	40	45	50	55	60	65	70	75	80	85
5	82.1	41.2	27.6	20.9	16.9	14.3	12.5	11.1	10.1	9.3	8.7	8.3	7.9	7.6	7.4	7.3	7.2
10	41.0	20.6	13.8	10.5	8.5	7.2	6.2	5.6	5.0	4.7	4.4	4.1	3.9	3.8	3.7	3.6	3.6
11	37.3	18.7	12.6	9.5	7.7	6.5	5.7	5.0	4.6	4.2	4.0	3.8	3.6	3.5	3.4	3.3	3.3
12	34.2	17.2	11.5	8.7	7.0	6.0	5.2	4.6	4.2	3.9	3.6	3.4	3.3	3.2	3.1	3.0	3.0
15	27.4	13.7	9.2	7.0	5.6	4.8	4.1	3.7	3.4	3.1	2.9	2.8	2.6	2.5	2.5	2.4	2.4
20	20.5	10.3	6.9	5.2	4.2	3.6	3.1	2.8	2.5	2.3	2.2	2.1	2.0	1.9	1.8	1.8	1.8
25	16.4	8.2	5.5	4.2	3.4	2.9	2.5	2.2	2.0	1.9	1.7	1.6	1.6	1.5	1.5	1.4	1.4
30	13.7	6.9	4.6	3.5	2.8	2.4	2.1	1.8	1.7	1.6	1.4	1.4	1.3	1.3	1.2	1.2	1.2
35	11.7	5.9	3.9	3.0	2.4	2.0	1.8	1.6	1.4	1.3	1.2	1.2	1.1	1.1	1.0	1.0	1.0
40	10.3	5.2	3.5	2.6	2.1	1.8	1.6	1.4	1.3	1.2	1.1	1.0	1.0	1.0	0.9	0.9	0.9
45	9.1	4.6	3.1	2.3	1.9	1.6	1.4	1.2	1.1	1.0	1.0	0.9	0.9	0.8	0.8	0.8	0.8
50	8.2	4.1	2.8	2.1	1.7	1.4	1.2	1.1	1.0	0.9	0.9	0.8	0.8	0.8	0.7	0.7	0.7
55	7.5	3.7	2.5	1.9	1.5	1.3	1.1	1.0	0.9	0.8	0.8	0.8	0.7	0.7	0.7	0.7	0.7
60	6.8	3.4	2.3	1.7	1.4	1.2	1.0	0.9	0.8	0.8	0.7	0.7	0.6	0.6	0.6	0.6	0.6
65	6.3	3.2	2.1	1.6	1.3	1.1	1.0	0.8	0.8	0.7	0.7	0.6	0.6	0.6	0.6	0.6	0.6
70	5.9	2.9	2.0	1.5	1.2	1.0	0.9	0.8	0.7	0.7	0.6	0.6	0.6	0.5	0.5	0.5	0.5
75	5.5	2.8	1.8	1.4	1.1	1.0	0.8	0.7	0.7	0.6	0.6	0.6	0.5	0.5	0.5	0.5	0.5
80	5.1	2.6	1.7	1.3	1.0	0.9	0.8	0.7	0.6	0.6	0.5	0.5	0.5	0.5	0.5	0.4	0.4
85	4.8	2.4	1.6	1.2	1.0	0.8	0.7	0.6	0.6	0.5	0.5	0.5	0.5	0.4	0.4	0.4	0.4
90	4.6	2.3	1.5	1.2	0.9	0.8	0.7	0.6	0.6	0.5	0.5	0.4	0.4	0.4	0.4	0.4	0.4
95	4.3	2.2	1.4	1.1	0.9	0.8	0.6	0.6	0.5	0.5	0.4	0.4	0.4	0.4	0.4	0.4	0.4
100	4.1	2.1	1.4	1.0	0.8	0.7	0.6	0.6	0.5	0.5	0.4	0.4	0.4	0.4	0.4	0.4	0.4

APPENDIX

PART K

Bellamy Gradient Wind Nomogram



APPENDIX

PART L

Spacing of isotherms by the shear vector speed at different levels.*

Spacing of isotherms at the 850 mb level at selected latitudes for $\Delta T = 2°$ C. The shear vector is between winds at 1200 m (4000 ft) and 1800 m (6000 ft) ASL. Shear wind speed is in kt and Δn is in ° Lat.

Lat \ kt	4	8	12	16	20	24	28	32	36	40	44
70	1.5	0.7	0.5	0.4	0.3	0.3	0.2	0.2	0.2	0.2	0.1
60	1.6	0.8	0.6	0.4	0.3	0.3	0.2	0.2	0.2	0.2	0.2
50	1.8	0.9	0.6	0.5	0.4	0.3	0.3	0.2	0.2	0.2	0.2
40	2.2	1.1	0.7	0.5	0.4	0.4	0.3	0.3	0.2	0.2	0.2
30	2.8	1.4	0.9	0.7	0.6	0.5	0.4	0.4	0.3	0.3	0.2
20	4.1	2.0	1.3	1.0	0.8	0.7	0.6	0.5	0.4	0.4	0.4
10	8.1	4.1	2.6	2.0	1.6	1.4	1.1	1.0	0.9	0.8	0.7

Spacing of isotherms at the 700 mb level at selected latitudes for $\Delta T = 4°$ C. The shear vector is between winds at 2450 m (8000 ft) and 3660 m (12000 ft) ASL. Shear wind speed is in kt and Δn is in ° Lat.

Lat \ kt	8	12	16	20	24	28	32	36	40	44	48
70	3.2	2.2	1.6	1.3	1.1	0.9	0.8	0.7	0.7	0.6	0.5
60	3.5	2.3	1.8	1.4	1.2	1.0	0.9	0.8	0.7	0.6	0.6
50	4.0	2.7	2.0	1.6	1.3	1.2	1.0	0.9	0.8	0.7	0.7
40	4.8	3.2	2.4	1.9	1.6	1.4	1.2	1.1	1.0	0.9	0.8
30	6.2	4.1	3.1	2.5	2.0	1.7	1.5	1.4	1.2	1.1	1.0
20	8.8	6.0	4.5	3.5	3.0	2.6	2.2	2.0	1.8	1.6	1.5
10	17.0	12.0	8.8	7.0	5.8	5.0	4.4	3.9	3.6	3.2	2.9

Spacing of isotherms at the 500 mb level at selected latitudes for $\Delta T = 4°$ C. The shear vector is between winds at 4800 m (16000 ft) and 6100 m (20000 ft) ASL. Shear wind speed is in kt and Δn is in ° Lat.

Lat \ kt	8	12	16	20	24	28	32	36	40	44	48
70	3.8	2.6	1.9	1.5	1.3	1.1	1.0	0.9	0.8	0.7	0.6
60	4.2	2.8	2.1	1.7	1.4	1.2	1.1	0.9	0.8	0.8	0.7
50	4.7	3.1	2.4	1.9	1.6	1.4	1.2	1.0	0.9	0.9	0.8
40	5.6	3.7	2.8	2.2	1.9	1.6	1.4	1.3	1.2	1.0	0.9
30	7.3	4.8	3.6	2.9	2.4	2.0	1.8	1.6	1.5	1.3	1.2
20	10.5	7.1	5.3	4.3	3.5	3.0	2.6	2.4	2.1	1.9	1.8
10	21.0	14.0	10.9	8.4	7.0	6.0	5.1	4.7	4.1	3.8	3.5

*When computing spacing at a given height the density of the air must be considered. The density and the equation of the thermal gradient becomes

$$\frac{dv}{dz} = \frac{g}{fT} \frac{\Delta T}{\Delta n} .$$

The values of ΔT were chosen to be 2° C and 4° C and Δn given in degrees latitude. The value of T in the Standard Atmosphere was used at each level. When the value of T is colder than the Standard Atmosphere, Δn would be larger and smaller when the air is warmer. However, this change is small since T is used in ° K. The standard level is not always in the middle of the layer because the height above mean sea level of the selected winds is constant but the height above mean sea level of the 850, 700, and 500 mb levels change.

APPENDIX

PART M

Specific use of upper air soundings.

Extracted from: NAVAER 50-1P-529, Handbook Single Station Analysis and Forecasting Techniques, January 1955.

The following pages were copied from AWSP 105-56 (1979a).

ABSTRACT: Lack of data, especially in wartime, will not lessen the demands for weather service. This article presents a technique that can be used to deduce the type of system that is influencing the local area. This technique requires only the latest upper air sounding and continuity (to construct an upper air sounding see AWSP 105-56, pp. 8.1.3-1). Knowledge of the type system present will allow the forecaster to prepare a better forecast.

I. ANTICYCLONIC CIRCULATIONS	PRESSURE	TEMPERATURE	LAPSE RATE	MOISTURE	TROPOPAUSE
1. Warm High Arriving.	Pressure rises at all levels in the troposphere rise increasing upward	Temperature rises in low and middle troposphere; falls above	Lapse rate stabilizes mostly in lowest levels, subsidence layers develop there	Moisture decreases through sounding	Rises, becomes indistinct toward center of upper high
2. Warm High Departing a Weak pressure gradient to west of high without active advance of MTW air	Pressure falls at all levels	No marked temperature change occurs in low and middle troposphere; slow warming above.	Stability decreases slowly; subsidence inversions disappear especially over land in summer.	Moisture increases gradually aloft over land in summer	Is very indistinct, often absent
b Active advance of MTW air to west of high.	Pressure fall begins. Maximum at ground and diminishing to about 6 kilometers; pressure rises above that point	Temperature rises in low and middle troposphere; falls in high troposphere and stratosphere.	See II, 1, (a)	Moisture increases sharply.	Rises markedly
3. Warm high increasing locally in intensity.	Pressure rises slowly at all levels	Temperature changes little in low and middle troposphere; cools slowly above.	Great stability with marked subsidence inversion.	Humidity is very low at all levels	
4. Warm High Decreasing Locally in Intensity.	Pressure falls slowly at all levels; fall increasing upward in troposphere	Temperature changes little in low and middle troposphere; rises slowly above	Stability intensifies somewhat over water, but slowly decreases over continental interiors in summer.	Upper air continues dry over water; moisture increases slowly over continental interiors in summer	Rarely recognizable
5. Cold High Subsiding and Changing Into Warm High.	Pressure is high at ground, steady, rising from about 3 kilometers upward; largest increases are in stratosphere	Temperature is below normal at ground initially, but rises between ground and about 6 kilometers with cooling in high troposphere and stratosphere.	Stability is extreme with several subsidence inversions in lower troposphere.	Air is very dry at all levels	Rises and becomes indistinct in central section of high
II. CYCLONIC CIRCULATIONS					
1. Cold Low Arriving. Outskirts. a If active warm front of tropical maritime air is at hand	Pressure fall begins, maximum at ground and diminishing to about 6 kilometers; pressure rises above that point	Temperature rises in low and middle troposphere; falls in high tropsophere and stratosphere.	Stable to frontal inversion. Inversion weak if MTW is unstable. **Above Inversion:** Unstable MTW indicated by lapse rate increasing to moist adiabatic	Moisture increases sharply, especially above front	Rises markedly
b For weak warm front with MP air.	Pressure fall begins in low and middle troposphere; maximum fall at ground little change in high layers	Moderate warming in low and middle troposphere	or greater with convection and cumulo type clouds. Stable MTW is indicated by continued or increasing stability with stratus type clouds. (Changes less marked for weak warm front.)	Some moisture increases, especially above front	Rises some; is frequently ill defined

CYCLONIC CIRCULATIONS, Cont.	PRESSURE	TEMPERATURE	LAPSE RATE	MOISTURE	TROPOPAUSE
2. Cold Low Arriving At Approach of Cold Front. a. Wind normal to front increases with height. Cold front moves more rapidly than warm air above it which ascends cold front slope and then moves eastward aloft ahead of front	Pressure fall intensifies at all levels	Temperature drops between 2.5 or 3 to 4 or 5 kilometers; rises below	Instability increases through troposphere	Moisture increases markedly aloft, near saturation; prefrontal middle and high clouds appear at various levels	**MTW Air In Warm Sector** Tropical tropopause rises to peak shortly before arrival of cold front. Arctic tropopause comes in rapidly underneath it, and tropical tropopause disappears soon. Cold tropical air in high troposphere is rapidly displaced by warm Arctic air and a low stratosphere. **NP air In warm sector** Arctic tropopause forms as before, but higher tropopause is very indistinct
b. Cold front moves more slowly than the air above it which is generally polar maritime and descends this cold front slope	Pressure fall intensifies at all levels	Temperature rises in lower troposphere; cools from about 5 kilometers upward in troposphere	Stability increases in lower levels with subsidence layers; marked instability from about 5 kilometers upward in troposphere	General drying occurs	
3. Cold Low Arriving: Near Upper Center.	Pressure fall is most intense at all upper levels, increasing upward through troposphere, small changes near ground	Temperature falls sharply through troposphere; rises in stratosphere	Troposphere is generally unstable; cold front inversion is very weak or absent.	Moisture is at saturation through sounding except in lowest layers, if polar continental air is advancing	Dips sharply; is at minimum elevation
4. Cold Low Departing	Pressure rises at all levels; maximum rise near ground.	Temperature rises through troposphere except in lowest layers in polar air; slow cooling stratosphere	Stability increases through sounding	Moisture constant but relative humidity decreases.	Rises and becomes indistinct
5. Cold Low Deepening.	Pressure in renewed fall, with largest decreases in high troposphere	Troposphere in renewed cooling, except at lower levels where warming occurs ahead of secondary cold fronts	Instability increases with prefrontal cloudiness	New increases in moisture and prefrontal cloudiness	New lowering of tropopause
6. Cold Low Filling; before passages of main trough; winds from invariant directions, decreasing.	Pressure fall lessens, or stops entirely.	Cooling of troposphere, except lowest layers in polar air, is slowed down or stops; similarly temperature rise in stratosphere is slowed down or even cooling may begin	Instability decreases through sounding	Moisture about constant, cloud decks thin out	Is steady and becomes less distinct; higher tropopause may form over it
7. Unstable polar front Waves.		USE STANDARD CYCLONE MODELS			
8. Warm Low Approaching	Pressure falls rapidly near ground; little pressure change at and above 3 kilometers	Temperature increases rapidly in low troposphere, rises somewhat in high troposphere	Small changes in stability	Minor moisture increases in lower troposphere	Absent or weak

APPENDIX

PART N

TARWI Code.

The code and specifications as given in AWSR 105-24 VOL one, March 1983.

Target Weather Reporting Code (TARWI).

a. This is a standard NATO code designed primarily for target weather reports from strike aircrews.

b. Form of code:

TARWI YQ$L_aL_aL_a$ $L_oL_oL_o$G'G' CHVWR

or

$X_BX_BX_B$ $X_LX_LX_NX_NX_NX_N$ YYGG CHVWR

c. Specification of Symbolic Letter(s):

TARWI	Code identifier
Y	Day of week (as given in WMO Code 4900)
Q	Octant of globe (as given in WMO Code 3300)
$L_aL_aL_a$	Latitude in tenths of degrees
$L_oL_oL_o$	Longitude in tenths of degrees
G'G'	Time of observation to the nearest quarter hour GMT (24 hour clock). G'G' gives times to nearest quarter hour. Time 15 minutes past hour -- add 25 to hour for GG Time 30 minutes past hour -- add 50 to hour for GG Time 45 minutes past hour -- add 75 to hour for GG
$X_BX_BX_B$	Identification of Grid Zone Designation
X_LX_L	100,000 meter square identification
$X_NX_NX_NX_N$	Numerical grid coordinates of the observation point given to the desired accuracy
YY	Day of month
GGgg	Time of observation in hours and minutes GMT
C	Low cloud amount in eighths. Coded 9 if not observed
H	Low cloud height
V	Visibility
W	Weather
R	Remarks

d. Specifications.

H — Low Cloud Height

0 -- No low cloud
1 -- 500 ft or less
2 -- 1000 ft
3 -- 1500 ft
4 -- 2000 ft
5 -- 2500 ft
6 -- 3000 ft
7 -- 3500 ft
8 -- 4000 ft
9 -- not observed

V — Visibility

Kilometers	Nautical Miles
0 -- 0 to less than 1	0 to less than ½
1 -- 1 to less than 2	½ to less than 1
2 -- 2 to less than 3	1 to less than 1½
3 -- 3 to less than 4	1½ to less than 2
4 -- 4 to less than 5	2 to less than 3
5 -- 5 to less than 6	3
6 -- 6 to less than 7	More than 3 to less than 4
7 -- 7 to less than 8	4 to less than 5
8 -- 8 or greater	5 or greater
9 -- V is not reported	V is not reported

W — Weather

0 -- not observed
1 -- no significant weather
2 -- sleet
3 -- dust or smoke
4 -- fog or haze
5 -- drizzle
6 -- rain
7 -- snow
8 -- showers
9 -- thunderstorms

R Remarks

A -- encoded weather is simulated (for exercise use)

B -- multiply cloud heights by 10

C -- no medium or scattered medium clouds

D -- scattered variable broken medium clouds

E -- broken variable overcast medium clouds

F -- contrails at flight level

G -- enroute weather predominantly IFR

H -- enroute weather predominantly VFR

I -- gusty winds at surface

J -- fog in valley

K -- higher terrain obscured

L -- surface conditions variable due to showers

M -- thunderstorms occurring

N -- thunderstorms enroute

O -- icing at flight level or freezing precipitation

P -- surface wind NE quadrant

Q -- surface wind SE quadrant

R -- surface wind SW quadrant

S -- surface wind NW quadrant

T -- weather better to north

U -- weather better to east

V -- weather better to south

W -- weather better to west

X -- weather suitable for mission

Y -- weather marginal for mission

Z -- weather unsuitable for mission

The remarks entry selected will be the one considered most significant for the mission.

APPENDIX

PART O

Short guidelines for operations in tropical rainfall areas - Henry (1972)

Introduction

Man's endeavor is often thwarted by the weather. In the tropics the meteorological variable is the rainfall. The temperature at a place has a daily range which usually exceeds the annual range, but the limits are fairly uniform and not exceeded to a critical point. The wind has a predominance from one quadrant, and the speed is not excessive except during the few rare occasions when a major tropical storm moves through the area. This leaves the precipitation as the major variable and the most important for the populace because the economy of most tropical areas is based on agriculture, and variations of the rains are all important. The tropical areas are subject to the extreme rainfall and the threat of flooding, as well as lack of rainfall and drought.

Some past concepts of tropical rainfall have expressed a sameness of rainfall by such sayings as "will meet for tea after the shower." Others view the rainy season as a faucet which is opened at the beginning of the season and closed at the end. Such concepts are totally in error.

The only statement which can be made with assurance is as follows: "The rainfall is extremely variable in both the areal and temporal dimensions." The rain last year, yesterday, and today has little in common with any other rainfall. It can be dry in the rainy season and rainy in the dry season.

The General Nature of Tropical Rainfall

Most of the tropical rainfall occurs in the form of rain showers. Each shower represents a mesoscale system. Showers are identified as having a bursting start, short duration (one h.), large drops, and being scattered about over the countryside, with each shower covering a small area. A shower may be accompanied by thunder and lightning. All these conditions indicate a conditionally stable atmosphere which may at any time or place become unstable and the mesoscale rainstorm develop.

On a larger scale (macroscale) the atmosphere takes on certain characteristics. Large scale cyclones and anticyclones develop and move

both at the surface and aloft. In the tropics the surface cyclones and anticyclones are not especially noticeable (except major storms). The change aloft is not great but is sufficient so that permissive or repressive periods occur. Permissive or repressive periods relate to the macro conditions which permit or inhibit the development of the mesoscale rain showers. During some periods (days to months in duration), the large scale conditions prohibit the mesoscale rain showers in an area. A few mesoscale systems may develop to a sufficient extent to cause a few isolated rain showers. This is why there are always a few rains during the dry season.

It is possible to observe the movement of the permissive atmosphere by observing where the showers are abundant. The movement is usually slow (<100 mi/day). The permissive area is too large to observe at one station but requires many stations over a large area.

Heavy rainfall amounts occur with tropical storms. Hurricanes cause copious rains over large areas. The rainfall, however, is caused by many mesoscale systems, closely bound together and organized by the hurricane. The rainfall is not uniform across an area but is varied by both amount and time of occurrence. At any one place the rainfall varies with intensity as the mesoscale rain showers pass over.

The "temporale" is another type of rainstorm which occurs in the tropical areas. This storm is caused by a permissive macro system, and the mesoscale systems are not very active. The storm appears more like the general area rain of the more northerly latitudes. There are a number of different definitions and descriptions of this rain storm. Each country has its own variations.

The light rain and drizzle of the northerly latitudes is not very common in the tropics. It does occur, but it occurs more frequently at higher elevations where the atmosphere is colder than at sea level. The amounts of rain and drizzle are small when compared with the amounts of water from the rain showers.

The general guidelines which follow apply to the usual condition, but will not take care of all cases and all areas. The most frequent occurrence is described and some of the exceptions noted.

General Guidelines for Daily Rainfall

1. The rainfall is greater 10-15 mi inland than at the coast.

This applies to a fairly flat littoral and not to coast lines that are mountainous. Some of the beaches are almost a desert, but much more rain falls inland. The cause is--the moist air which is flowing onshore needs to be heated by the land surface to cause instability and the showers.

2. In the tropics a heavy rain at one point is an indicator that little or no rain is occurring 20 mi away. The chance for rain about 35 mi away is better than at 20 mi when it is raining at your location.

This rule is best on flat lowland areas; and in general, applies to single meso system rainfall. When air is rising within the cell, there is subsidence for a distance around the cloud. The cloud diameters are about 15 mi; so that distance 20 mi away is outside of the cloud. By a distance of 35 mi the next cell has formed and rain occurs. This is illustrated by Fig. 0-1. In mountainous areas the cells are confined and do not necessarily grow to these sizes. Also, the terrain dictates where much of the rain will fall.

On the higher plateau area such as Savanna de Bogota (Colombia), the cold air does not hold as much moisture; so the size of the cells is less and the distances reduced. When major systems (cold fronts, hurricanes, etc.) pass over the area, the rain falls over the entire area but not necessarily uniformly.

3. Over most land areas only one rainstorm a day will occur.

A rainstorm for this purpose is defined as at least 10 mm (0.4 in). This rule verifies during about 90% of the rain days. Only a few places will have two rainstorms a day. However, Sihanoukville, in Cambodia, had seven rainstorms in one day. The rule is more applicable during the absence of major systems.

4. Most of the rain days are days of light rain.

There are many days on which only a little rain falls, but which are counted as rainy days. They do not contribute much to the monthly total. This is represented by Fig. 0-2, which shows Martin's universal curve. This curve is representative of most places. In places where the rain is great (40 in/mo), the curve would be to the upper left of Martin's curve.

5. A meso rainstorm will provide an inch of rain.

This rainfall is under the center of the storm. The fringe areas receive less. The center of the storm may provide more than one inch.

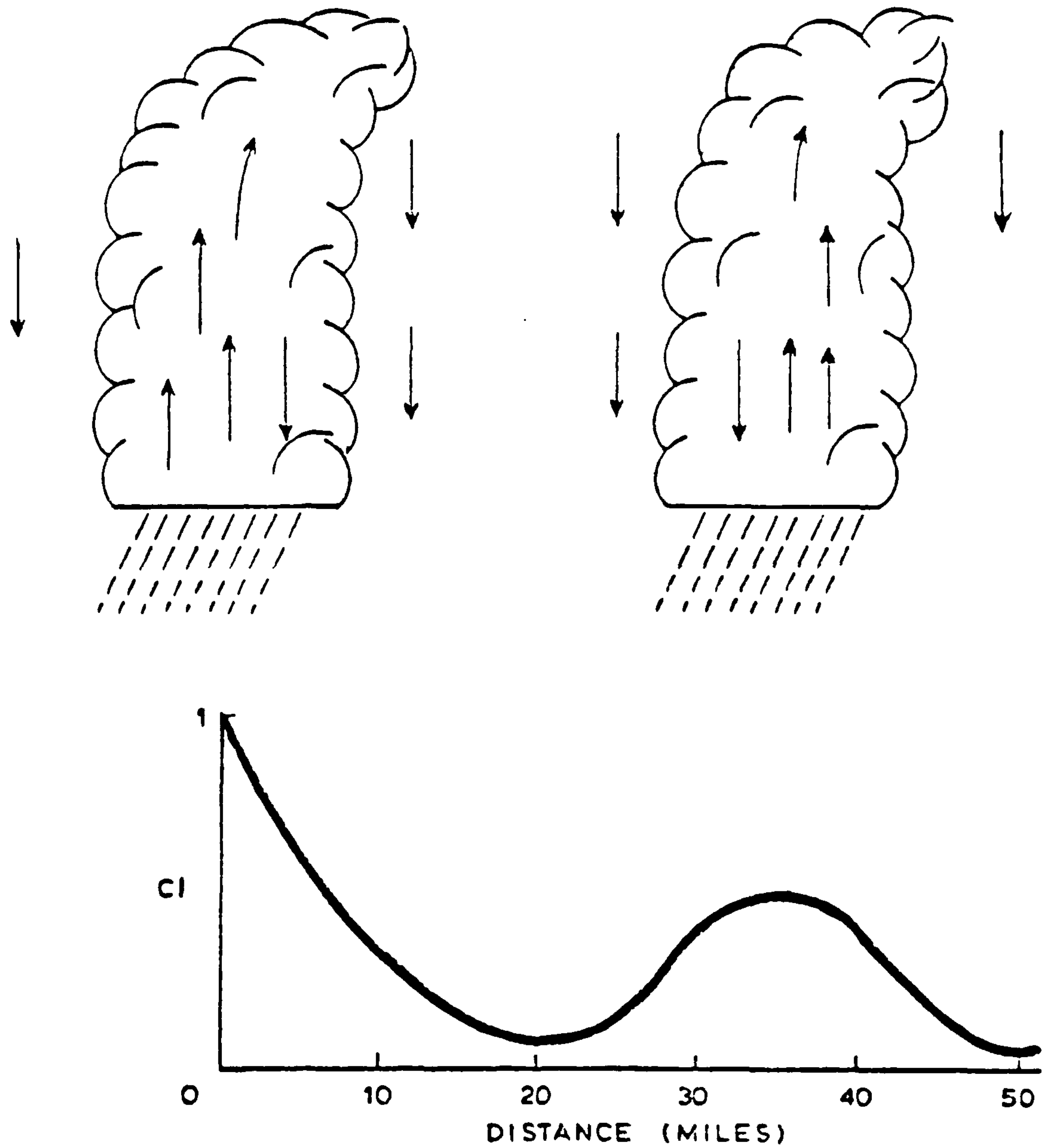


FIG. 0-1 Schematic curve of Contingency Index (CI) with distance. Mesoscale clouds and circulations are added to show the physical relationship.

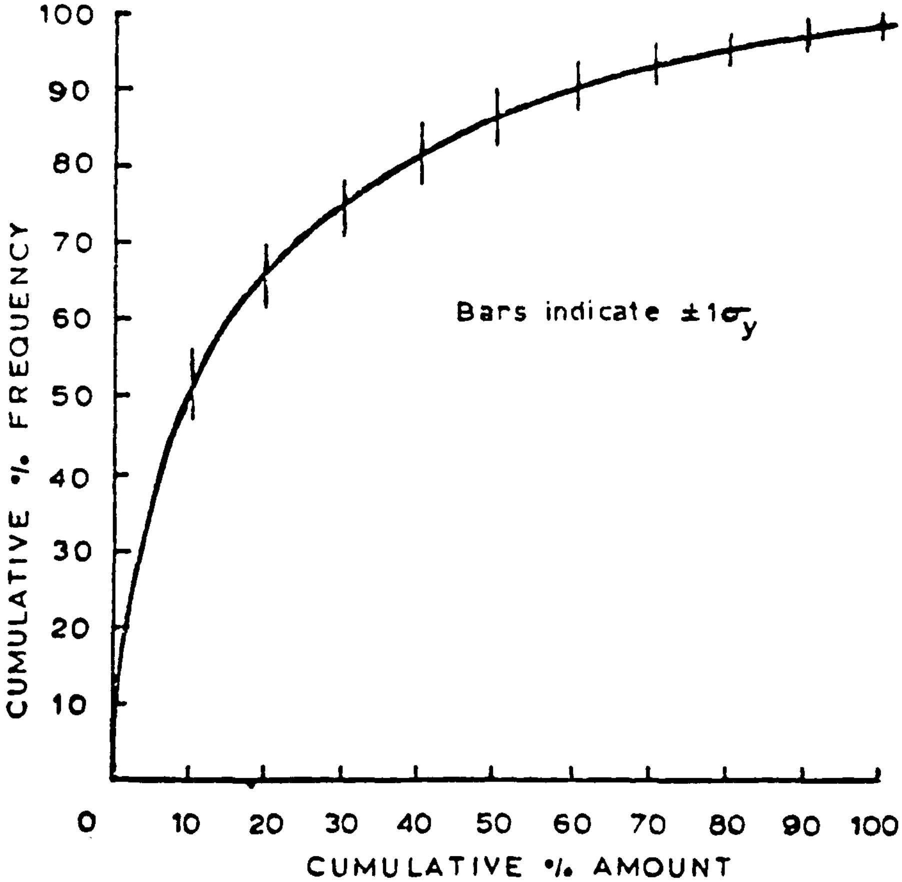


FIG. 0-2 Mean ogive of cumulative frequency vs cumulative amounts, both in per cent of total. The bars indicate a standard deviation to each side of the curve.

Two or three inches are common, and as much as eight inches is reported infrequently. The area which receives an inch or more is usually elliptical and about 2 mi across and 5 mi long. The fringe area extends 10 to 12 mi from the center.

Sometimes light rain does fall from fragmented mesoscale systems which do not have the core of heavy rain. Also, some light rain occurs from orographic lifting. This rainfall will not have the center core described in the previous paragraph.

The rainstorm does not start at certain clock times. Therefore, hourly rainfall can be a little misleading because a storm may start during the latter part of an hour and extend into the following hour. If the hour of heaviest rainfall is selected as the time to center the rainstorm upon, then the average Cambodian rainstorm will be as shown in Table 0-I.

TABLE 0-I

Hourly Rainfall Before and After the Hour of Maximum Rainfall (mm) in Cambodia During the Southwest Monsoon

Case	Hourly Total Before and After Maximum Hour						Storm Total	Daily Total
	-2	-1	Max	+1	+2	+3		
I	0	0	20.7	0	0.0	0	20.7	20.7
II	0	0	47.0	14.9	3.0	2.2	67.1	67.8
III	0	10.0	10.5	0	0	0	20.5	20.5
IV	2.5	13.2	36.6	7.7	8.9	4.6	73.5	74.0
V	0	2.6	36.2	33.5	.2	0	72.5	79.7
VI	5.0	11.5	37.5	0	0	0	49.5	49.5

When many individual cases as shown in Table 0-I are averaged together, the composite storm then would be as shown in Table 0-II.

TABLE O-II

Composite Mesoscale Rainstorms for Cambodia
During the Southwest Monsoon (inches)

Hourly Total Before and After Maximum Hour					average Storm Total
-2	-1	Max	+1	+2	
.03	.10	.65	.12	.04	.94

6. Rainstorms occur any time of the day. However, the morning hours (8 a.m. to noon) have the lowest occurrence. The maximum for most tropical areas is in the afternoon and early evening. Early morning (3 to 6 a.m.) also is common. In some places there is a seasonal variation of the day and nighttime rains.

7. Rain today - rain tomorrow.

Analysis of the data does not indicate that this is the case. During the rainy season it does rain most of the days, but days of no rain do occur. During the dry season, two or even three days of rain occur, usually in sequence.

General Guidelines for Monthly Rainfall

1. The rainy season starts and ends with regularity in the mean. When five-day totals of rainfall are used as an indicator of the beginning of the rainy season, a sharp change of the amount and frequency of rainfall can be observed. Over large areas of the tropics, a five-day total of one inch marks the onset of the rainy season. This is not an infallible guide because some areas do not receive this much rainfall. An example of the advance and retreat of the rainy season is shown in Fig. 0-3a and 0-3b. In the areas that receive less rainfall, the change of amounts can be checked for changes which will indicate that the rainy season has arrived or departed.

2. Last month's rainfall is not an indicator of next month's

rainfall. There is, in some areas and by some persons, a misconception that the rain of the previous month foretells the rainfall of the next month or even for the year. This concept is incorrect. There is no relationship of the rainfall of one month to the next nor of that month last year.

Also, there is a belief that the first three or four months of the year indicate the rainfall for the year. This does have a statistical

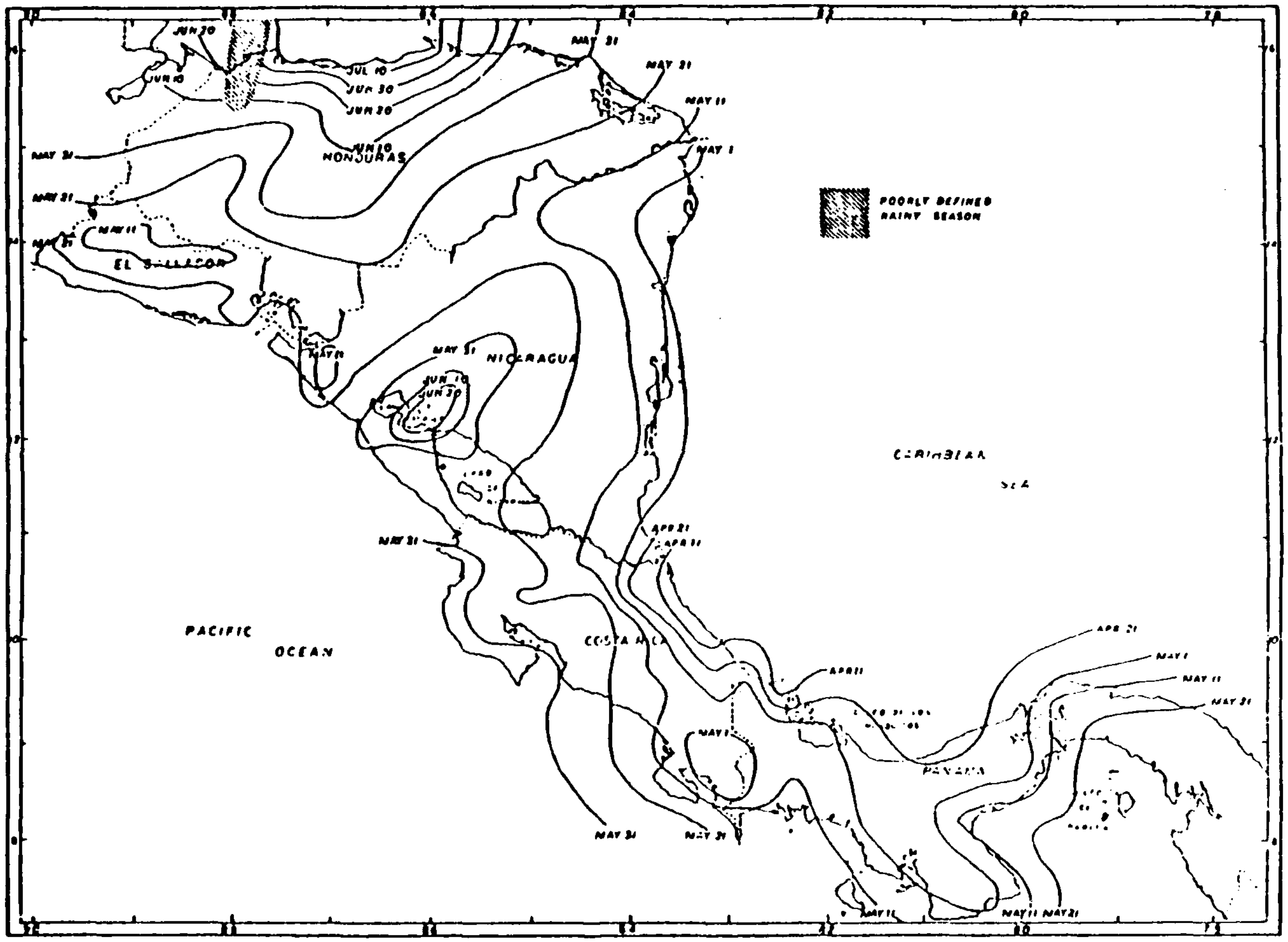


FIG. 0-3a Date of arrival of 50 percent chance of 25 mm rainfall in five days.

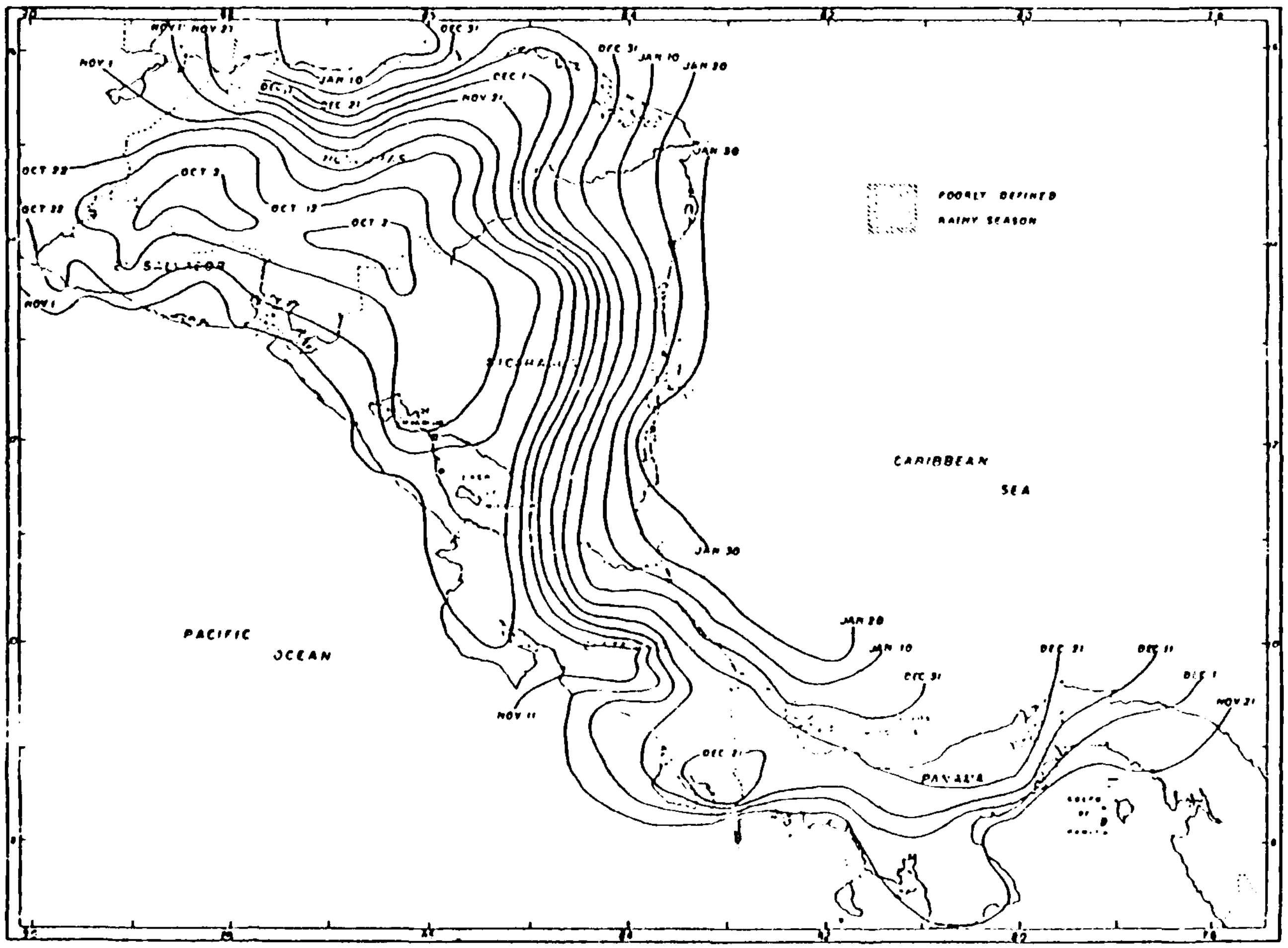


FIG. 0-3b Date of ending of 50 percent chance of 25 mm of rainfall in five days.

backing, but only because the first three or four months are part of the annual total.

3. Relationship between the monthly rainfalls of the two stations.

Two stations, further than 10 mi. apart and with no orographic features separating them, will not have the same monthly totals. Also, they will not necessarily have the same deviation from the mean of each station. They often deviate in the opposite direction. One will exceed its mean, while the station 10 mi. away will receive less than its mean. An example of the variation of the coefficient of association with distance is shown for Guyana in Fig. 0-4.

"Monthly rainfall is variable" is the only correct statement which can be made.

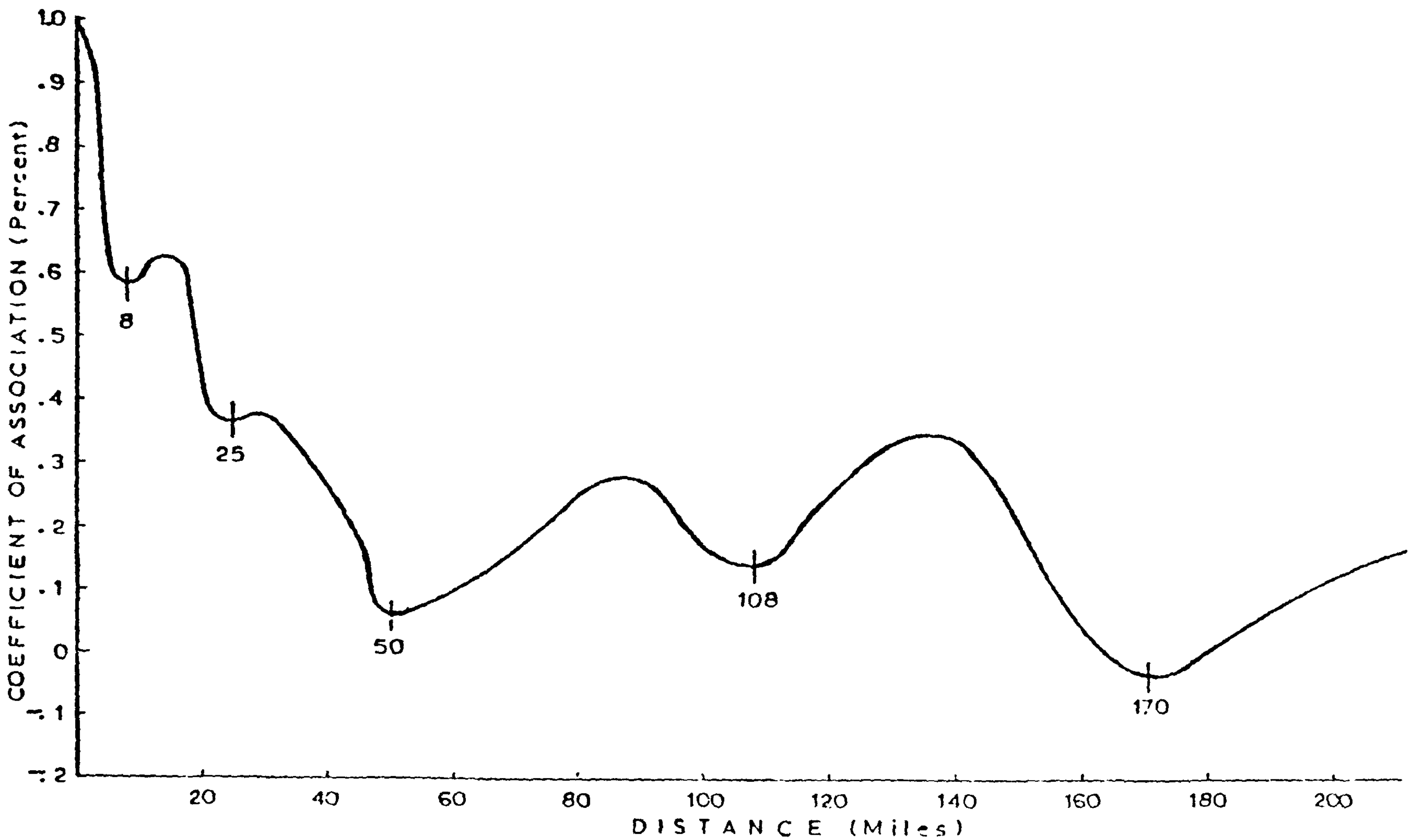


FIG. 0-4 Profile of the coefficient of association from Georgetown for the month of October.